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Quaternary environments, climate and man in the Netherlands

Field Guide

Edited by C. Kasse

Quaternary Research Association Annual Field Meeting, The
Netherlands, April 6-9th 1995, Vrije Universiteit, Amsterdam



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Climatic change and fluvial evolution of the Maas during the late Weichselian and early Holocene

QRA-excursion Saturday 8 April 1995

Climatic change and fluvial evolution of the Maas during the late Weichselian and early Holocene

C. Kasse

Introduction

The late Weichselian and Holocene evolution of the Maas (or Meuse) valley in northern Limburg, north of Venlo, is controlled by tectonic and climatic factors. Changes in climate, vegetation and river discharge resulted in changes in the fluvial depositional environment and in terrace formation.

The Maas river takes its rise at approximately 400 m above sea level in the Mesozoic rocks of the Paris Basin in eastern France. In its course to the north it cuts through the Paleozoic rocks of the Ardennes Massif (up to 700 m high) in Belgium. North of Maastricht it enters the southern North Sea Basin (fig. 1).

The Maas catchment is 33.000 km². The mean annual July temperature in the catchment area is between 15 and 18° C. The mean annual temperature in January is approximately between 0 and 2.5° C. There is little snowfall in winter. Mean annual snow coverage varies from less than 10 days at the coast to 35 days inland. Hence the Maas is a rain-fed river. The mean annual precipitation amounts 700 to 1000 mm and up to 1300 mm in the highest parts of the Ardennes. The maximum discharge is in January and the minimum discharge between July and September, but interannual variation in the discharge is very large (Jongman, 1987).

In the following table the mean, maximum and minimum discharges in m³/sec are given of the rain-fed Maas at Borgharen (Netherlands-Belgian border) and of the meltwater-fed Rhine at Lobith (Netherlands-German border) (Jongman, 1987).

| | Maas | Rhine |
|--------------------------|------------------------|-------------------------|
| period | 1911-1960 | 1901-1975 |
| catchment | 33.000 km ² | 185.000 km ² |
| mean summer discharge | 130 | 1850 |
| minimum summer discharge | 0 | 640 |
| maximum summer discharge | 1150 | 7150 |
| mean winter discharge | 390 | 2540 |
| minimum winter discharge | 0 | 620 |
| maximum winter discharge | 2800 | 13000 |

These figures clearly show the larger difference between the mean winter and mean summer discharge ($Q_{\text{mean winter}}/Q_{\text{mean summer}}$) of the Maas in comparison with the Rhine. Furthermore, the fluctuations in discharge of the Maas during the winter ($Q_{\text{max winter}}/Q_{\text{mean winter}}$) and especially during the summer ($Q_{\text{max summer}}/Q_{\text{mean summer}}$) are larger than those of the Rhine, illustrating the rain-fed character of the Maas.

In Decembre 1993-January 1994 and in January 1995 the Maas valley was

flooded by the highest floods ever recorded. At Borgharen discharges of 3120 m³/sec and 2870 m³/sec were measured during the two floods. The floods were caused by exceptionally high rainfall in the Ardennes and northern France.

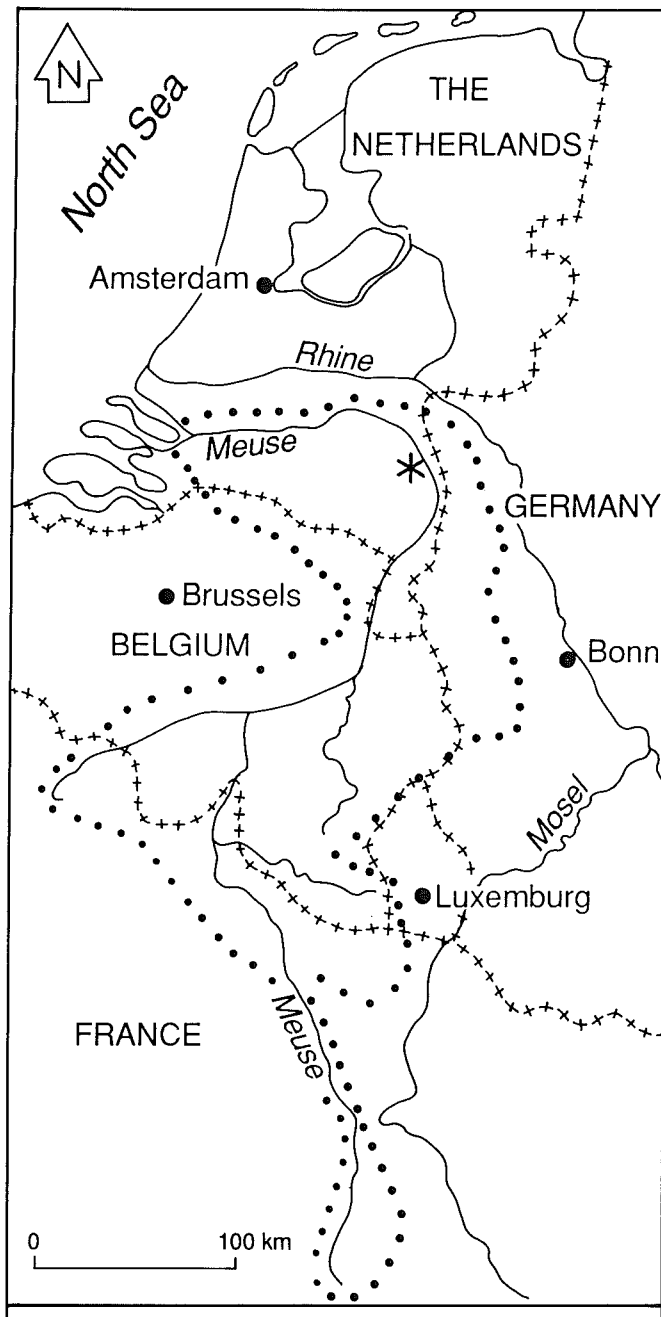


Fig. 1 Catchment area of the Maas (from Kasse et al., 1994)

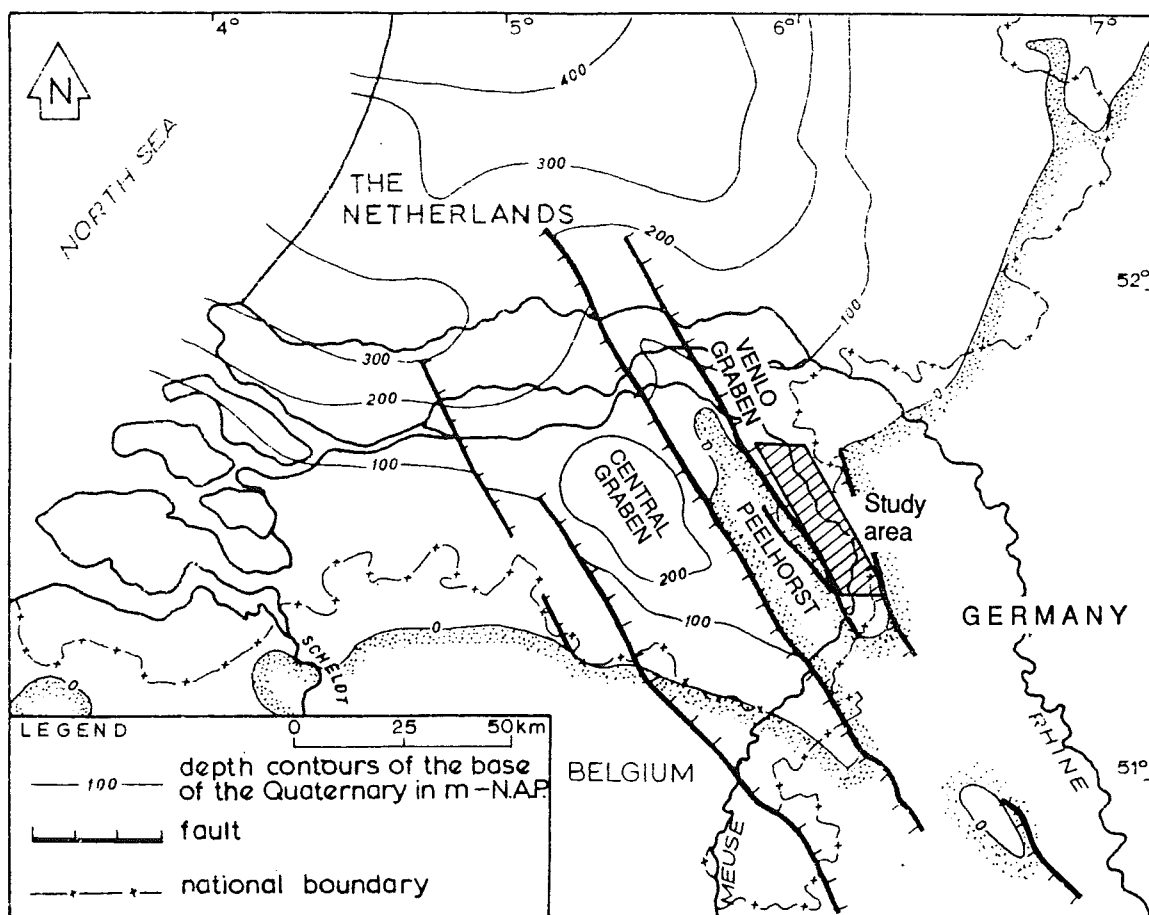


Fig. 2 Major tectonic units of the Southern North Sea basin and depth contours of the base of the Quaternary deposits (after Zagwijn & Doppert, 1978).

Tectonic setting

North of Maastricht the Maas enters the southeastern part of the North Sea Basin, which is characterized by the occurrence of southeast-northwest oriented faults (fig. 2). This fault system, which is the northwestern continuation of the Lower Rhine Graben, forms structural lows (Central Graben and Venlo Graben) and highs (Peel Horst) in the excursion area.

The Grabens contain a thick Quaternary sequence (up to 200 m in the Central Graben), which indicates the continuous subsidence during the Quaternary (Zagwijn, 1989). During the Middle Pleistocene (Cromerian) the Rhine and to a lesser extent the Maas occupied the Central Graben (Sterksel Formation). Due to a strong uplift of the Rhenish Plateau during the Late Cromerian (400.000 years ago), the Rhine changed its course to the north and formed the augite bearing Urk Formation (Zagwijn, 1989). In the Central Graben the Rhine was replaced by the Maas, which

deposited the Veghel Formation. During the Elsterian, Holsteinian and Saalian the Maas gradually shifted eastwards over the Peel Horst area into the Venlo Graben (Van den Toorn, 1967; Zagwijn & Van Staalduinen, 1975).

At present the Maas crosses the Central Graben and the Peel Horst almost at right angles, before it bends to the northwest in the Venlo Graben. The actual river morphology reflects the tectonic movements. In the Central Graben the Maas has a strongly meandering course with a broad floodplain (Van den Broek & Maarleveld, 1963). On the Peel Horst the Holocene floodplain is nearly absent along the straight, incised course. In the Venlo Graben the present river has a narrow floodplain and a low sinuosity meandering course.

Late Pleniglacial river evolution

During the glacial maximum of the Late Pleniglacial the Maas was a braided river system with large discharge fluctuations (fig. 3: levels 1 and 2; Excursion stop Grubbenvorst sand pit). Syngenetic ice-wedge casts point to permafrost conditions in the floodplain and a mean annual temperature below -8 °C (fig. 4). Aggradation prevailed because of the high sediment supply in the unvegetated landscape. During annual high water stages the braidplain was flooded and medium to coarse sand was deposited by transverse bars. In the last stage of each flooding event small-scale erosive channels were formed. Locally, stagnant pools developed in which silt was draped over the inactive bars. During a next high water stage the channels were filled laterally, with large-scale cross-bedded sand and climbing ripple cross-laminated sand, due to the migration of transverse bars over the floodplain.

Towards the end of the Late Pleniglacial, fluvial deposition became less important probably because of a stronger aridity in combination with a slight temperature increase. Parts of the braided plain became covered with aeolian sands (so-called coversands) and reworked aeolian sands (fluvio-aeolian facies of Grubbenvorst sand pit). The river maintained its braided character and flowed at more or less the same level as during the previous period (fig. 3: level 2).

Late Glacial river development and terrace morphology

Above the recent Holocene floodplain, four distinct Late Glacial to Early Holocene terrace levels have been distinguished along the Maas north of Venlo (fig. 3: levels 3, 4, 5, 6).

The transition of the Pleniglacial to the Late Glacial was accompanied by a temperature rise (fig. 4). The previously unvegetated Late Pleniglacial landscape was stabilized by vegetation. As a consequence the channel morphology of the Maas changed from braided into meandering and the river began to incise (Kasse et al., 1994; Vandenberghe et al., 1994) (figs. 4 and 5). In the abandoned braided channels peat formation started during the Bølling, therefore dating the moment of change in the fluvial environment (Bohncke et al., 1993; Excursion stop Bosscherheide).

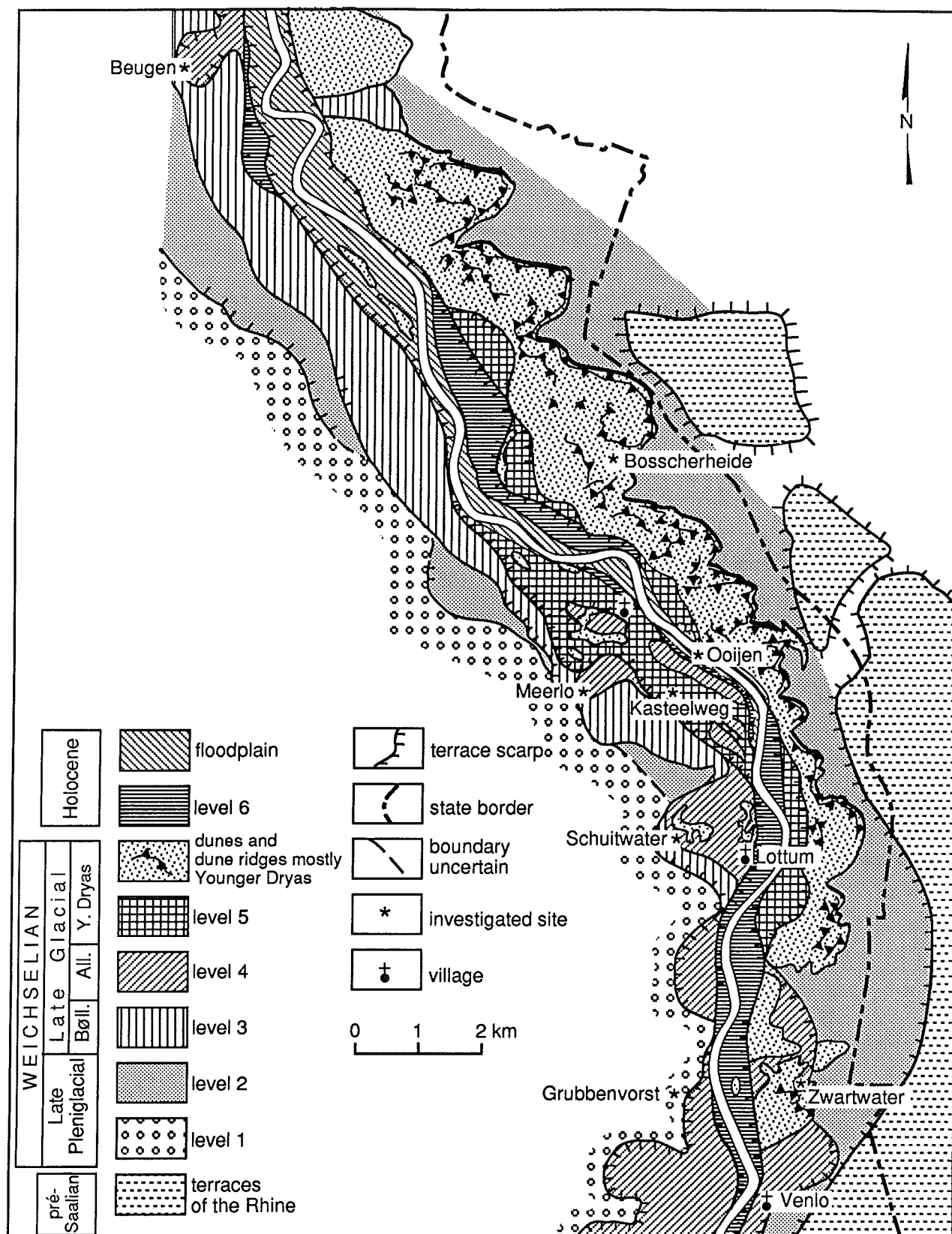


Fig. 3 Morphological map of the Late Pleniglacial, Late Glacial and Holocene terraces of the Maas north of Venlo (from Kasse et al., 1994, partly after Wolfert & De Lange, 1990).

This alteration from braided into meandering is characterized by a transitional phase with rather shallow (2-3 m), slightly incised, low-sinuosity channels (fig. 3: level 3). This transitional phase probably took place during or after the Bølling, but before the Allerød.

The next younger Late Glacial terrace is typified by large, high-sinuosity meander scars, especially at the outer terrace edges (fig. 3: level 4; Excursion stops Meerlo and Beugen). This terrace level descends from 21 m above sea level at Venlo to 13 m at Beugen. The fine-grained scar fills are 3 to 5 m thick. The organic fills in the meander scars, dating from the Allerød and Younger Dryas, indicate that this high-sinuosity meandering phase occurred during the Allerød (figs. 4 and 5). Individual pointbars are poorly developed because of an aeolian cover or intense human occupation and cultivation.

The sedimentary sequence of the meander scar terrace is characterized by a thick (7.5 m) fining-upward sequence, formed by lateral migration of the channel and accretion on the meander pointbar (fig. 3: Lottum-Schuitwater). The fining-upward sequence consists of 2 m gravelly, poorly sorted, medium to coarse sand (300-850 μm) at the base overlain by a transitional bed of circa 1 m of moderately sorted, fine to medium sand (150-300 μm). This coarse grained lower part was formed by strong tractional currents on the meander channel bottom and the lower part of the pointbar slope. The upper 4.5 m of the fining-up sequence are moderately or more often well sorted fine sands (105-210 μm) with thin sandy silt beds (1-13 cm), which increase in number and thickness towards the top. Some smaller fining-up sequences, separated by erosional boundaries, are present within this fine sand unit, probably reflecting reactivation surfaces of the meander inner bend during high discharges. The fine-grained, well sorted upper part was deposited by weaker tractional currents on the upper pointbar slope. The silt beds reflect slack water conditions following high water levels on the upper pointbar slope. These fines have been deposited from suspension by weak currents or in standing water.

The lowest Late Glacial terrace dates from the Younger Dryas (fig. 3: level 5). At the start of the Younger Dryas a strong temperature decline occurred (fig. 4), leading to local permafrost conditions. At Excursion stop Bosscherheide permafrost degradation was dated between 10,880 BP and 10,500 BP (Bohncke et al., 1993). Lower evapotranspiration and larger discharge fluctuations resulted in local flooding of the Late Pleniglacial braidplain at Bosscherheide. The high sinuosity meandering channels of the Allerød period were abandoned by chute cut offs and a braided river floodplain was formed. The higher discharges in combination with a rather restricted sediment supply resulted in erosion of the floodplain and an up to 4 m high terrace scarp was formed. The Younger Dryas floodplain declines in altitude from c. 18 m near Venlo to c. 10 m near Beugen (fig. 3: level 5). It is characterized by its straight to low sinuous scars (fig. 3: Excursion stop Kasteelweg). Straight scars occur especially along the terrace edge to the higher meander scar terrace. Islands or bars in this floodplain are locally covered by river dune sand, which was blown from the multi-channel plain during periodic low water levels. The lithology of the Younger Dryas terrace level reveals laterally a larger heterogeneity than the higher meander

terrace level 4. The palaeochannels are shallow and broad and locally contain up to 2 m of fine-grained infilling in isolated scour pits (Excursion stop Kasteelweg). They are underlain by coarse sand and gravel. The bars in between the channels consist of (gravelly) sand; fining-upward sequences being less pronounced and normally shorter than in the meander scar terrace.

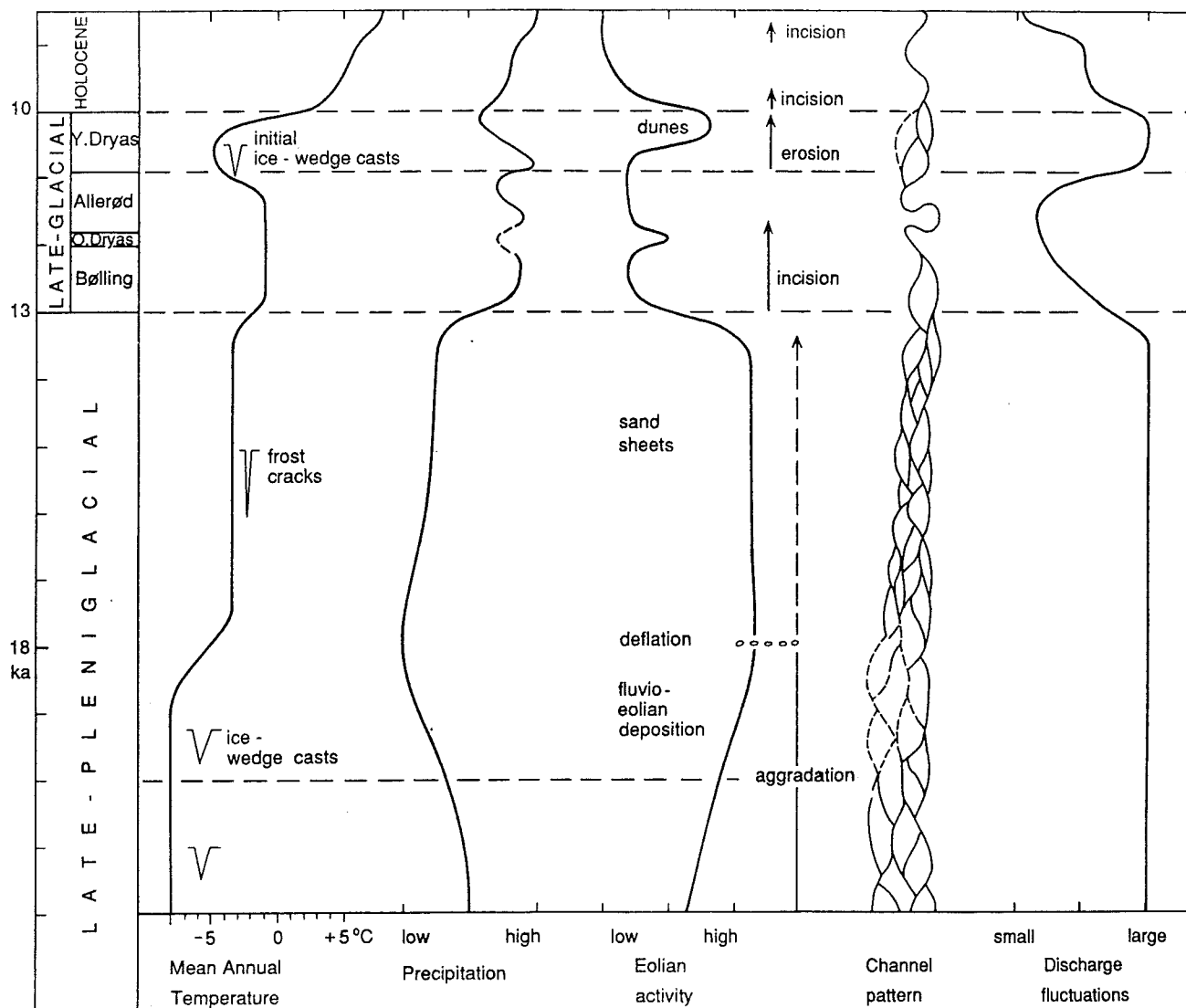


Fig. 4 Summary of the Late Pleniglacial and Late Glacial climatic changes and climate-related fluvial development of the Maas river (from Kasse et al., 1994).

The climatic amelioration at the start of the Holocene led to a higher land surface stability and to a decrease in the discharge fluctuations, resulting in river incision (figs. 4 and 5). The braided channels of Younger Dryas age were abandoned

and the river changed into a low-sinuosity meandering system during the Preboreal (fig. 3: level 6). The oldest fill in the incised channels dates from the Boreal (Excursion stop Ooijen), which indicates that the incision phase occurred between the Younger Dryas and the Boreal, i.e. the Preboreal. The Preboreal floodplain reveals straight and low-sinuosity scars, more or less conform to the actual low-sinuosity river course. Due to the lateral migration of the meander belt, channel side bars or large-scale pointbars developed. This clear fluvial morphology indicates that the aeolian activity, typical of the late Younger Dryas, had ceased.

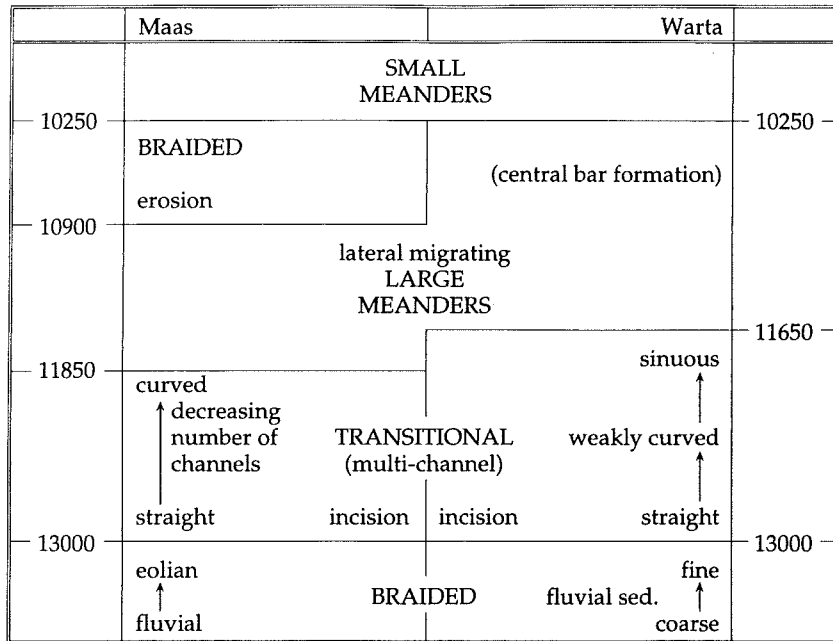


Fig. 5 Comparison of the Late Pleniglacial and Late Glacial climate-related fluvial developments of the Maas (the Netherlands) and Warta (Poland) rivers (from Vandenberghe et al., 1994).

Late Pleniglacial and Late Glacial aeolian activity

During the very end of the Late Pleniglacial aeolian deposition was the dominant process over large areas in the Netherlands. This widespread aeolian deposition can be attributed to aridity due to the absence of permafrost and/or lower precipitation values. In the Maas valley Late Pleniglacial fluvio-aeolian and aeolian deposits have been described by Mol et al. (1993) (see Excursion stop Grubbenvorst).

Late Glacial aeolian deposits are also very common in the Maas valley. Aeolian sediments dating from the Allerød and/or Older Dryas period are only locally present on the higher meander terrace level (fig. 3: level 4, Lottum-Schuitwater). At this site the meander terrace is covered by an up to 4 m thick unit of aeolian sediments. Since there is no evidence for an aeolian supply from outside the area, nor for a supply

from the subsoil, we conclude that these aeolian sediments on top of the meander terrace were formed during the lateral migration of the meandering river channel. The comparable fine sandy grain-size of the aeolian deposits and the pointbar deposits support this hypothesis. The following mechanism has to be considered. During bankfull discharge, probably in spring due to melting of the snow cover, sediment was deposited on the meander upper pointbar. During the following low discharge this barren sediment on the upper slope could be deflated by the prevailing westerly winds (Maarleveld, 1960; Schwan, 1988) and was deposited on top of the point bar. The low dune morphology indicates aeolian deposition in a vegetated landscape. The start of the meander scar infill at Lottum-Schuitwater was dated palynologically in the Younger Dryas. This means that the lateral channel migration and the connected deflation and dune accumulation took place just before the Younger Dryas, i.e. the Allerød and/or Older Dryas period.

The younger, far more extensive, Late Glacial aeolian deposits date from the Younger Dryas. On the east side of the Maas valley widespread river dune complexes occur, lying on the Late Glacial meander scar terrace or on older fluvial deposits (fig. 3: Excursion stop Bosscherheide). The dune morphology is characterized by parabolic forms in the eastern (downwind) part of the dune field. Because of this morphology a westsouthwestern sand-transporting wind is inferred for the Younger Dryas period. It is stressed here that this wind direction is the dominant sand-transporting wind, probably generated by the passage of low-pressure systems with associated high wind velocities. The dominant annual wind direction may have been completely different. At the base of the dune sediments Late Glacial organic sediments and peats are present, which offer the opportunity to date the start of the aeolian deposition. From the top of the peaty layer, characterized by an alternation of moss-laminae and aeolian sand-laminae, a ^{14}C date of $10,500 \pm 60$ BP was obtained (Bohncke et al., 1993), which places the overlying dune body in the late Younger Dryas period.

The aeolian sediment of this phase is fine to medium grained. Because of the prevailing westsouthwesterly winds, the source area must have been the Younger Dryas floodplain west of the dune field. It is proposed that the aeolian sediment has been blown from the Younger Dryas floodplain during periodic low water levels. Only the sediments that accumulated outside the floodplain have been preserved. The aeolian sediment within the floodplain had a low preservation potential because it was easily eroded during subsequent high water floods.

The accumulation of large amounts of sand in the extensive dune field on the east bank of the Maas could only occur due to the repeated fluvial reworking of the braided floodplain followed by deflation. These alternating processes of reworking and deflation indicate that during the late Younger Dryas the Maas had a more intermittent character than during the previous Late Glacial periods. This is in agreement with the morphological evidence from the floodplain, which shows a braided river system during this period. The larger fluctuations in discharge are attributed to the climatic deterioration at the start of the Younger Dryas period. The larger volumes of snow melt water will have led to larger peak discharges. As a result

the Late Glacial meandering river course changed into the braided Younger Dryas course.

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Late Glacial and early Holocene vegetational and climatic history in the Netherlands

S. Bohncke

The Late Glacial vegetation development of the Netherlands and North-West Europe has recently been summarized by Bohncke (1993) and Walker et al. (1994) (figs. 6 and 7).

The earliest accumulation of organic matter under favourable palaeohydrological conditions seems to have taken place shortly after 13 Ka BP. An AMS-date from the type-locality of Usselo (Van Geel et al., 1989) provided a date of 12,930 BP. At 12,885 BP organic sedimentation started in the former Dinkel Valley (Ran, 1990). Conventional ^{14}C -dates, based on larger quantities of organic matter, indicate an onset of organic production in abandoned shallow river channels at around 12,700-12,600 BP (Bohncke et al., 1987; Teunissen, 1983).

Prior to 13 Ka BP a shortage in effective precipitation may well have been the limiting factor for the spread of shrubs and trees (e.g. Van Campo, 1984). In the absence of a reasonable vegetation cover the surface was prone to wind activity and a continuous transport of sand led to the formation of coversands (Older Coversands II, Van der Hammen, 1971). The Older Coversands II consist of horizontally bedded fine sands. Sedimentary structures that indicate running water are rare and periglacial features are absent except for some faint micro drop-soil structures. Pollen analyses of the oldest infill of pingo remnants (e.g. Mekelermeer; Bohncke et al., 1988a) indicate the presence of a heliophilous herbaceous vegetation. *Helianthemum*, *Rumex acetosa/acetosella*, *Polemonium*, *Thalictrum*, *Artemisia* and some *Chenopodiaceae* and *Plantago major/media* form important constituents of this unique vegetation.

12,600-12,000 BP: the Bølling biozone

With an increase in effective precipitation shortly after 13 Ka BP a *Betula nana* (dwarf birch) and *Salix* spp. shrub tundra expands and *Artemisia* becomes more frequent. The substratum gradually became fixed and aeolian activity faded out. The presence of freely drained soils in combination with an increase in precipitation favoured the spread of *Juniperus* (c. 12,600 BP) which often forms a distinct vegetation belt before tree birches shade these out.

With the establishment of tree birches the AP/NAP ratio increases and a boreal birch forest seems to have been the dominant vegetation type. Biostratigraphically this pollen assemblage zone (*Betula-Salix-Artemisia* p.a.z.) is correlated with the Bølling.

Fig. 6
Summarized Late Glacial
vegetation succession
for the Netherlands
(from Bohncke et al., 1988).

| HOLOCENE | P.A.Z. | CHRONOSTRATIGRAPHY | VEGETATION CHARACTERISTICS | SOIL CONDITION |
|----------|---|----------------------|---|---|
| | 10,000 | | | |
| | <i>Betula</i> , <i>Empetrum</i> , <i>Cyperaceae</i> | YOUNGER DRYAS | Expansion of a sedge-rich <i>Empetrum</i> heath, intermingled with <i>Betula</i> and <i>Salix</i> dwarf shrubs. Tree birches and pine trees are sparse, but more prominent in the S. Netherlands. | Soil conditions instable. Deposition of cover sands. Leaching and acidification prevails. |
| | 11,000 | | | |
| | <i>Betula</i> / <i>Pinus</i> | | | |
| | <i>Pinus</i> / <i>Betula</i> | ALLERÖD | Some <i>Empetrum</i> heath development with <i>Juniperus</i> . Boreal <i>Pinus</i> forest expansion, with <i>Populus tremula</i> ; <i>Filipendula</i> increases. | Soil formation and start of podsolisation process. |
| | <i>Betula</i> (Pinus) | | | |
| | 11,800 | | | |
| | <i>Salix</i> , <i>Betula</i> , <i>Artemisia</i> . | OLDER DRYAS | Regional shrub vegetation replaced by some <i>Juniperus</i> and <i>Hippophae</i> shrubs. In favourable conditions tree birches maintain. | Soil conditions instable, soil erosion and deposition of coversands |
| | 12,000 | | | |
| | <i>Betula</i> , <i>Salix</i> , <i>Artemisia</i> . | BÖLLING | Shrub vegetation with <i>Betula nana</i> , <i>Salix</i> spp. and some <i>Juniperus</i> . Tree birches gradually spread. | Development of a vegetation cover and start of soilforming processes. |
| | 12,400 | | | |
| | <i>Betula nana</i> , <i>Cyperaceae</i> , <i>Poaceae</i> . | OLDEST DRYAS | Heliophilous herbaceous vegetation with steppe tundra elements, intermingled with dwarf willow and dwarf birches. | Instable soil conditions. Lime-rich soils. |
| | 13,000 | | | |

| ¹⁴ ka BP | | IRELAND | ENGLAND, WALES & S. SCOTLAND | SCOTLAND HIGHLANDS & ISLANDS | NORTH BELGIUM | NETHERLANDS | N.W. GERMANY |
|---------------------|--|--|--|--|--|--|--|
| 9.5 | | WOODLAND <i>Corylus</i> <i>Betula</i> | WOODLAND <i>Corylus</i> <i>Betula</i> | OPEN WOODLAND <i>Betula</i> <i>Juniperus</i> <i>Empetrum</i> | <i>Pinus-Corylus</i> woodland | <i>Betula-Pinus</i> | WOODLAND <i>Corylus</i> <i>Betula pendula</i> woods |
| 10.0 | | OPEN WOODLAND <i>Betula-Juniperus</i> <i>Juniperus</i> OPEN SCRUB | TRANSITION TO WOODLAND <i>Juniperus</i> | TRANSITION TO OPEN WOODLAND | <i>Pinus-Betula</i> woodland | BOREAL FOREST <i>Pinus-Betula-Populus</i> | with <i>Populus tremula</i> and <i>Pinus</i> <i>Betula pubescens</i> woods |
| 10.5 | | <i>Artemisia</i> <i>Rumex</i> Gramineae | <i>Artemisia-Rumex</i> | <i>Artemisia-Rumex</i> | <i>Betula-Pinus</i> open woodland | <i>Juniperus-Betula nana</i> <i>Pinus</i> in S. <i>Empetrum</i> in N. <i>Betula-Salix</i> | <i>Juniperus</i> copses park Tundra with <i>Betula</i> , <i>Empetrum</i> & |
| 11.0 | | TUNDRA & LOW ALPINE SCRUB <i>Cruclerae</i> | TUNDRA & LOW ALPINE SCRUB | TUNDRA & LOW ALPINE SCRUB Gramineae-Cyperaceae Caryophyllaceae Lycopodium | <i>Pinus-Betula</i> open woods with <i>Empetrum</i> | <i>Empetrum-Artemisia</i> | <i>Betula pubescens</i> in N. & W.; <i>Betula-Pinus</i> woodland in SE |
| 11.5 | | Gramineae-Helianthemum | <i>Pinus</i> (SE England) <i>Betula</i> woods (Sporadic in N) | <i>Empetrum</i> & <i>Erica</i> Heathland | <i>Pinus-Betula</i> woodland <i>Betula</i> woods | <i>Pinus</i> Boreal woodland <i>Betula</i> Boreal woods | <i>Betula pubescens</i> woods with <i>Populus tremula</i> & <i>Pinus</i> in S. |
| 12.0 | | OPEN GRASSLAND | (Extensive in S) WOODLAND | Scattered <i>Betula</i> wood & scrubland | TRANSITION TO WOODLAND OPEN GROUND COMMUNITIES | TRANSITION TO WOODLAND <i>Salix-Betula-Juniperus-Hippophae</i> OPEN SCRUBLAND | TRANSITION TO WOODLAND <i>Juniperus</i> copses HERBACEOUS COMMUNITIES |
| 12.5 | | <i>Juniperus</i> or <i>Juniperus-Empetrum</i> (Depending on soils) | <i>Betula-Juniperus</i> wood & scrub | <i>Betula-Juniperus</i> scrub | Open <i>Betula</i> woods <i>Juniperus</i> | <i>Betula-Salix-Juniperus</i> | <i>Betula nana</i> , <i>Juniperus</i> , <i>Hippophae</i> , <i>Salix</i> (edaphically-controlled) |
| 13.0 | | LOW SCRUB & HEATHLAND COMMUNITIES <i>Rumex-Salix</i> <i>Artemisia-Gramineae</i> OPEN GROUND COMMUNITIES | <i>Juniperus</i> <i>Salix</i> OPEN WOOD & SCRUB | <i>Juniperus</i> OPEN WOOD & SCRUB HEATHS (N & W) | <i>Salix-Betula</i> TRANSITION TO WOOD & SCRUB OPEN GROUND COMMUNITIES | TRANSITION TO WOOD & SCRUB <i>Gramineae-Artemisia-Cyperaceae</i> <i>Gramineae-Helianthemum</i> <i>Thalictrum</i> <i>Chenopodiaceae</i> TUNDRA COMMUNITIES | TRANSITION TO SCRUB <i>Artemisia</i> STEPPE COMMUNITIES |

Fig. 7
Summary of the vegetation history during the Late Glacial in northwest Europe (from Walker et al., 1994).

12,100-11,800 BP: the **Early Dryas** biozone

The **Early Dryas** biozone, which is defined as a temporary decline in the AP and a concomitant increase in the NAP and which is recognizable in Late Glacial lake sediments, does not show up in the pollen assemblages of terrestrial peat sequences. Two possibilities can account for this discrepancy 1) terrestrial organic deposits underwent a period of non registration and 2) site inherent factors causing the same time-stratigraphic event to be registered differently in the pollen assemblages. In both cases the palaeohydrological conditions of the sites involved play an important part. The palaeobotanical record suggests a dry continental climatic episode for this time interval. The lowering of the phreatic level during the Early Dryas is reflected in the pollenrecord by a regional decline in the *Betula* boreal woodland. A return to a shrub and heliophilous herb vegetation is registered (the *Salix-Juniperus-Artemisia-Gramineae* p.a.z.).

In terrestrial sites, where the hydrosere succession is already more towards telmatic and semi-terrestrial, conditions will react to this relatively dry period either with a next stage in the vegetational succession, being the development of a local birch carr on top of the preceding stage with sedges and pleurocarpe mosses or will respond with hiatus(es) to this period with pronounced decline in effective precipitation.

Soil instability and the inwash of minerogenic sediments, which is reflected in the geochemical record of lake sediments (Bohncke & Wijmstra, 1988), characterize this zone. Terrestrial sites show the influx of aeolian sediments trapped in the local vegetation. A date of $11,990 \pm 70$ BP is available to pinpoint this event (e.g. site Mariahout; Bohncke, 1993).

11,900-10,900 BP: the **Allerød** biozone

Palaeohydrological studies reveal the return to high lake-levels (Bohncke & Wijmstra, 1988) for the start of this period. The increase in the phreatic level determines the termination of a period with possible hiatuses. Locally, conditions became suitable for aquatic species to participate in the hydrosere succession (Bohncke et al., 1993). The pollenrecord demonstrates a regional spread of tree birches in the southern Netherlands while at more northern latitudes this *Betula* increase is preceded by a *Juniperus* maximum. Both taxa rely on increased amounts of precipitation.

At c. 11,300 BP sites adjacent to river systems register a second phase with wetter conditions and occasionally the deposition of loamy lenses in their sequence is evident. An increase in (fluvial) erosion is inferred from this (Bohncke et al., 1987; Bohncke et al., 1993). The local hydrosere succession is set back to open water. In the subsequent local hydrosere succession the rhizomatose perennials like *Typha* spp. and *Phragmites* are lacking (e.g. Excursion site Bosscherheide, Bohncke et al., 1993) which leads to the supposition that winter temperatures must have dropped

and the annual temperature range augmented. Consequently, a more intensive action of the freeze-thaw cycle started to operate and instable soil conditions re-established.

Moreover, the lacustrine environments at this moment in time show an increase in *Isoëtes* sp., *Elatine hydropiper*, *E. hexandra* (Bohncke & Wijnstra, 1988). Both species require pioneer conditions. *Isoëtes* as a lake bottom dweller and *Elatine* spp. on the borders of the lake. It is not unlikely that longterm pioneer conditions were created by ice action.

Remarkable in this respect is the almost simultaneous westward spread of *Pinus*. Both in Usselo (Van Geel et al., 1989) and in Bosscherheide (Bohncke et al., 1993) this increase is dated to 11,300 BP. Since *Pinus* can be regarded as a more continental species, with its plant geographical centre of distribution at that period located in eastern Europe, it is concluded that the two features are somehow related. The spread of *Pinus* in this context must be interpreted as a response to climatic changes at around 11,300 BP implying more severe winter temperatures and increase in freeze-thaw incidences. The latter process is responsible for the increase in instable soil conditions and possible the increase in sediment load in the fluvial systems as reflected in the deposition of overbank deposits.

The termination of the **Allerød**, both in lake and terrestrial peat sequences, shows relatively dry conditions. Both a decline in effective precipitation and an increased evapotranspiration may be responsible for this phenomenon.

10,900-10,500 BP: the **Late Dryas** biozone

At c. 10,850 BP both vegetation and coleoptera (Bohncke et al., 1987) indicate a distinct drop in the average July temperature from between 18 and 15 °C to 11 and 10 °C. Consequently, the mean annual temperatures declined (estimated between -2 and -5 °C) and conditions approached those of permafrost environments. Simultaneously effective precipitation increased considerably resulting in a pronounced rise in the lake-levels (the lacustrine environment) and large-scale floodings and deposition of suspension load in large shallow lakes in the fluvial environment. Reconstruction of the vegetational history over the Allerød - Late Dryas transition quite often is obscured by erosion hiatuses and redeposition.

Evaluation of the available data in the Netherlands, taken into account the limitations set out above, provides a picture of a rapidly declining *Pinus* boreal forest. The pine forest was replaced by an open shrub vegetation with *Betula (nana)* and *Salix* spp. (e.g. *Salix reticulata*). Especially the presence of dwarf willows, generally known to be chinophilous, may indicate that at least a large part of the precipitation fell as snow. Heliophylous herbs became relatively more frequent (*Artemisia*, *Thalictrum*, Cyperaceae, Chenopodiaceae, *Helianthemum* and in some cases *Polemonium*). The local vegetation is characterized by the increase in aquatic taxa such as *Potamogeton* spp., *Batrachium* spp., *Hippuris vulgaris*, *Myriophyllum spicatum*, *Myriophyllum alterniflorum*, *Menyanthes trifoliata*, *Equisetum palustris* and *Sparganium* spp. and pleurocarpe mosses. This period came to an end when

precipitation diminished and the previously established large shallow lakes dried out at c. 10,500 BP.

10,500-10,200 BP: the **Late Dryas** biozone

Soon after, c. 10,500 BP, summer temperatures appear to have risen while effective precipitation declined considerably resulting in a decline in the lake-levels and periodically emerging river beds. Aeolian activity prevailed building up large parabolic river dunes (Bohncke et al., 1993).

At around 10,500 BP permafrost had disappeared completely and mean annual temperatures increased to about -1 °C. With the disappearance of the discontinuous permafrost the phreatic level declined, not to be restored by sufficient precipitation. In the absence of a protecting snow cover conditions may have been even more harsh for plant-life. This is emphasized in the pollen assemblages of this period by the presence, although sparse, of *Ephedra* spp. and *Polemonium*. The vegetational response to this dry upper part of the **Late Dryas** differs considerably, depending on the geographical position of the study-site.

In the northern part of the Netherlands this episode is preeminently characterized by *Empetrum* heath intermingled with dwarf birches and dwarf willows. Locally, some tree birches and pine trees may have survived. The NAP values are relatively low and dominated by sedges.

In the southern Netherlands *Pinus* spreads to reach dominance over *Betula* (tree birch and dwarf birch) by c. 10,400 BP (Vandenberghe & Bohncke, 1985).

In the SE Netherlands both *Betula* and *Pinus* persisted in equally high percentages intermingled with some *Juniperus* and *Betula nana* shrubs. The herb vegetation is dominated by grasses and *Artemisia*. Locally, *Helianthemum* and *Polemonium* may become important species in the sequence.

The prevailing dry conditions on the other hand frequently led to the hiatus(es) in the sequences, embracing the later part of the Late Dryas and the early Holocene.

10,200-9,100 BP: the **Preboreal** biozone

The Late Dryas period comes to an end when precipitation restores leading to an expansion of *Betula pubescens* s.l. In the northern and southeastern Netherlands this birch maximum is preceded by a *Juniperus* maximum. In the southern Netherlands, where *Pinus* forms the major tree during the later part of the Late Dryas, *Betula* increases and sometimes an interval in which *Betula* and *Pinus* reach equally high values is registered. This first episode of climatic amelioration of the early Preboreal is called the **Fiesland-phase** (10,200-9,850 BP).

Subsequently more continental conditions involving a decline in effective precipitation and possibly also in winter temperatures, determines the vegetational development: the **Rammelbeek-phase** (9,850-9,750 BP). Again tree-birches (*Betula*

pubescens s.l.) suffer from a decline in the phreatic level in the upland regions. An increase in grasses and a rapid hydrosere succession is registered at the site of the Borchert, northern Netherlands (Van Geel et al., 1981). Towards the end of this phase *Populus tremula* becomes a pioneer species on the emerging fringes of the mires. In the area within the pine-forest limits of that time, the more continental conditions result in a further expansion of *Pinus*, while the areas adjacent to these limits register their first Holocene *Pinus* expansion ($9,705 \pm 55$ BP, Schuitwater Lottum, Maas valley). At many localities the dry conditions again may result in a period of non-registration.

During the **Late-Preboreal** (9,750-9,150 BP) *Betula* restores from its temporary low values during the Rammelbeek-phase. An increase in the lake-levels for this period (Bohncke, 1991) indicates a restoration of the effective precipitation, which may account for the registered *Betula* increase. This time *Betula verrucosa* is involved in the vegetational succession (Van Geel et al., 1981).

In the southern Netherlands, where *Pinus* dominates the picture, small-scale fluctuations in the *Pinus* pollen values are registered. The dips in the *Pinus* curve consistently seem to coincide with an increase in local wet conditions (aquatics, *Salix*) and a slight increase in the *Betula* values. At 9,400 BP *Pinus* spreads in the northern part of the Netherlands, some 1000 C-14 years later than the *Pinus* increase in the SE Netherlands. *Pteridium aquilinum* and *Melampyrum* form part of the understory of the pine forests that dominated the late Preboreal vegetation cover.

The record demonstrates that the *Pinus* immigration and expansion is not simply a gradual northwards spread during the early Holocene but that superimposed on this process, periods of swift expansion designate the step-like expansion of the species.

9,150-8,000 BP: the **Boreal** biozone

The start of this zone is characterized by a gradual increase in *Corylus* (Hazel). *Corylus* is a heliophilous species and its spread during the Boreal implies a rather open forest for this period. Almost simultaneously *Quercus* (Oak) and *Ulmus* (Elm) appear in the vegetational record in the S. Netherlands accompanied by low values for *Tilia* (Lime). Towards c. 8,400 BP *Pinus* generally shows a maximum before it started to decline during the later part of the Boreal. Although lake-levels are generally low during the whole of the Boreal, this *Pinus* maximum coincides with a minimum in the lake-levels.

From c. 8,400 the forest canopy becomes more dense. *Pinus* values decline, while *Corylus* and the Quercetum mixtum (*Quercus*, *Ulmus*, *Tilia* and some *Fraxinus* (Ash)) expands. Climbers like *Hedera helix* and *Viscum album* occur in low frequencies, indicating that winter temperatures from now on are less severe (mean January temperature higher than -2°C). Towards the end of the Boreal a last *Pinus* maximum is often evident in the pollen record.

8,000-5,000 BP: the **Atlantic** biozone

The Atlantic period is characterized by a fast expansion of *Alnus*, which quickly outranges the *Pinus* values of the Late Boreal. The rapid increase in *Alnus*, a species that requires rather wet conditions, implies major environmental changes. It is assumed that air circulation patterns have changed dramatically and led to an increase in (effective) precipitation. This process will have initiated a rise in the phreatic level and a spread of water logged sites, the habitat that *Alnus* needs for its successful spread in the early Atlantic period. This is in agreement with the distinct rise in lake-levels found for the early Atlantic period (Bohncke, 1991). Between 7,000 and 6,700 BP *Fraxinus* reaches its rational limit (a continuous curve).

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Saturday 8th April 1995

Excursion program

Stop 1: GRUBBENVORST SAND PIT

- * Late Pleniglacial fluvio-aeolian facies
- * Late Pleniglacial cryogenic structures (ice-wedge casts, deformations, vertical platy structures)
- * Late Pleniglacial and Late Glacial aeolian sediments

Stop 2: BOSSCHERHEIDE GRAVEL PIT

- * Late Pleniglacial braided floodplain
- * Bølling to Allerød peat in abandoned channel
- * Younger Dryas embryonal ice-wedge casts and cryoturbations
- * Younger Dryas flooding and moss peat growth
- * Younger Dryas sand sheet and river dune formation

Stop 3: MEERLO CHANNEL

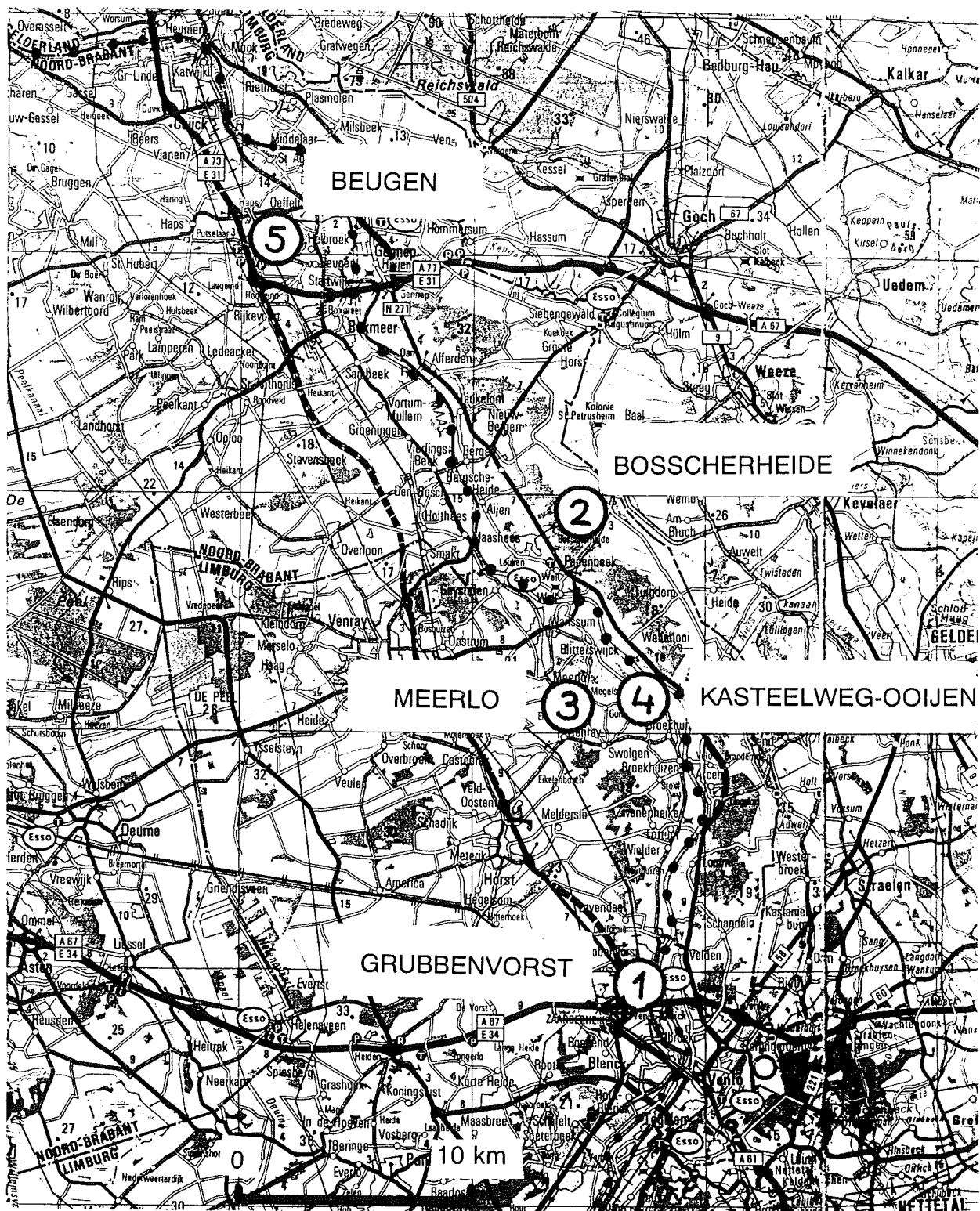
- * Allerød meander scar
- * Younger Dryas clastic infill

Stop 4: KASTEELWEG - OOIJEN CHANNELS

- * Younger Dryas braided floodplain morphology
- * Late Younger Dryas and Preboreal fill
- * Younger Dryas dune complex
- * Early Holocene incision and terrace edge
- * Boreal and Atlantic channel fill

Stop 5: BEUGEN

- * Late Glacial (Allerød) meander neck cut off
- * Allerød organic fill (gyttja)
- * Younger Dryas reactivation and clastic deposition



Location map with the excursion stops in the Maas valley.

EXCURSION STOP 1: GRUBBENVORST

Weichselian Late Pleniglacial fluvio-aeolian deposits and cryogenic structures

after J. Schwan & J. Vandenberghe (1991)

Stratigraphy

The Grubbenvorst sandpit is situated in a Maas river terrace that is overlain by a sheet of aeolian coversand (Fig. 1). The stratigraphical position of the exposed section (see Figs. 2 and 3) can be deduced from the following data:

1. The occurrence, some four km northwest of the exposure, of a subsurface peat layer that has been dated at GrN-16949 : $26,130 \pm 180$ BP at its base and at GrN-16948 : $25,200 \pm 180$ BP in its top (Westerhoff & Broertjes, 1990). This bed is presumed to be time-correlatable with a level found at a shallow depth below the floor of the Grubbenvorst sandpit.

2. The presence, in the top of unit B3 of the Beuningen Complex. At the site under consideration, this important marker bed consists of multiple deflation levels which formed during a protracted phase of intense drought. Following Kolstrup (1980) the development of this feature took place, roughly, between 15.5 and 14 ky BP.

3. The presence of a strongly iron-stained level in the upper part of the exposure. This feature supposedly represents a palaeosol of Alleröd Interstadial age and as such should have formed in the timespan from 11.8 to 11 ky BP.

On the basis of the above evidence, a Late Weichselian Pleniglacial to Weichselian Late Glacial age may be attributed to the section under consideration.

Fluvial to aeolian sedimentary facies

The Grubbenvorst section is a fine specimen of a fluvial to aeolian sequence, a type of succession that is wide spread in the Pleistocene lowlands of northwestern Europe (Schwan, 1987). In the exposure a gradual transition from a purely fluvial sediment at the base to a purely aeolian one in the top of the section is present (Figs. 2 and 3).

Unit A

In the fluvial sands of unit A, in the basal part of the section, large-size infilled channels are the prevailing structural characteristic. The channels occur in both intersecting as well as solitary positions, mostly have a concordant infilling and are incised in horizontally bedded sands. In upward direction, the channel fills give way to generally thinner sets which, on average, are also finer-grained, so that the whole of unit A represents a sequence with fining-upward and thinning-upward tendency.

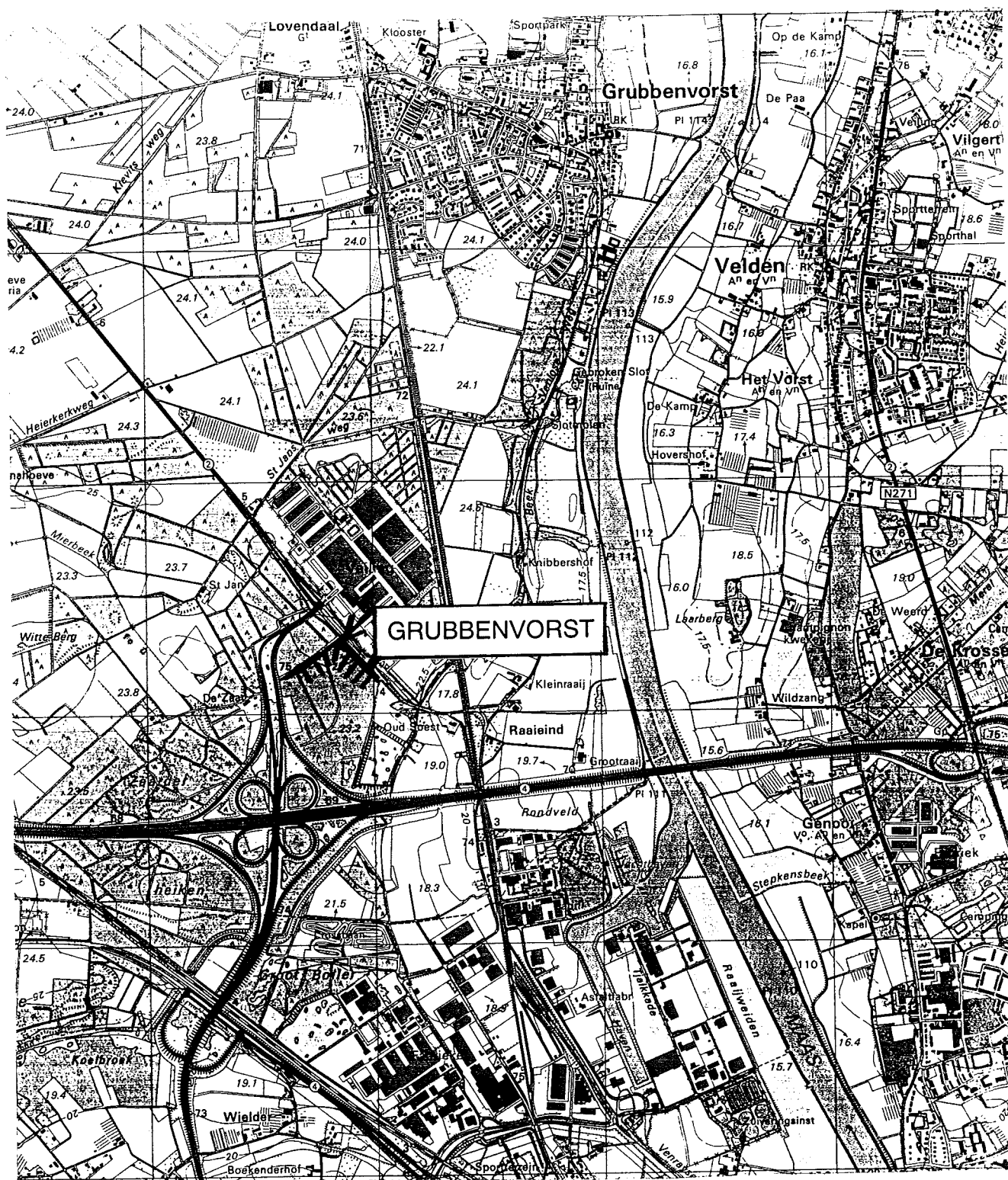
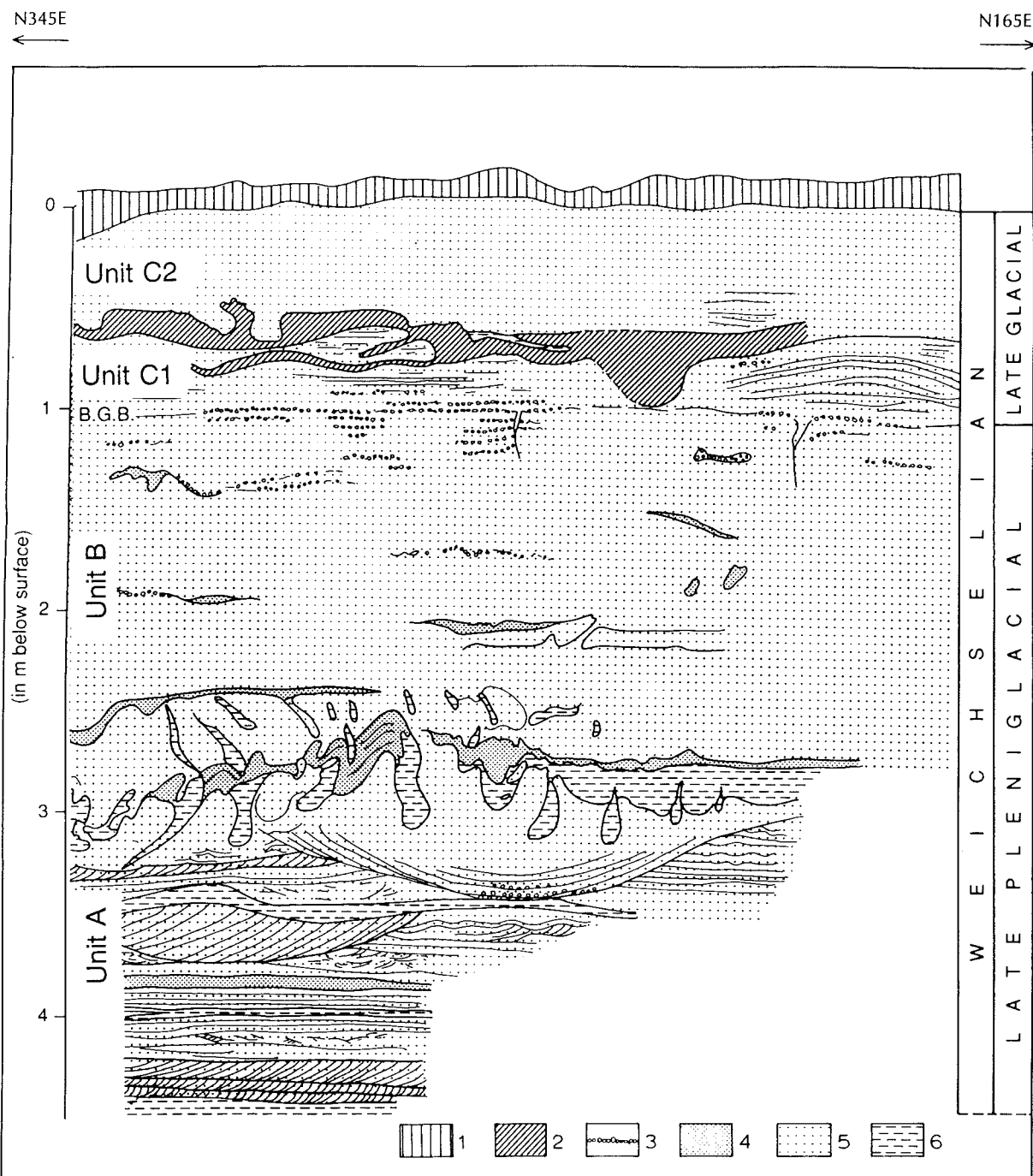


Fig. 1 Location map of the Grubbenvorst sand pit.



1, disturbed; 2, firm horizon; 3, gravel; 4, coarse sand; 5, very fine to medium sand; 6, silt; B.G.B., Beuningen Gravel Bed.

Fig. 2 Sedimentary succession in sand pit Grubbenvorst in July 1989 (taken from Mol et al., 1993).

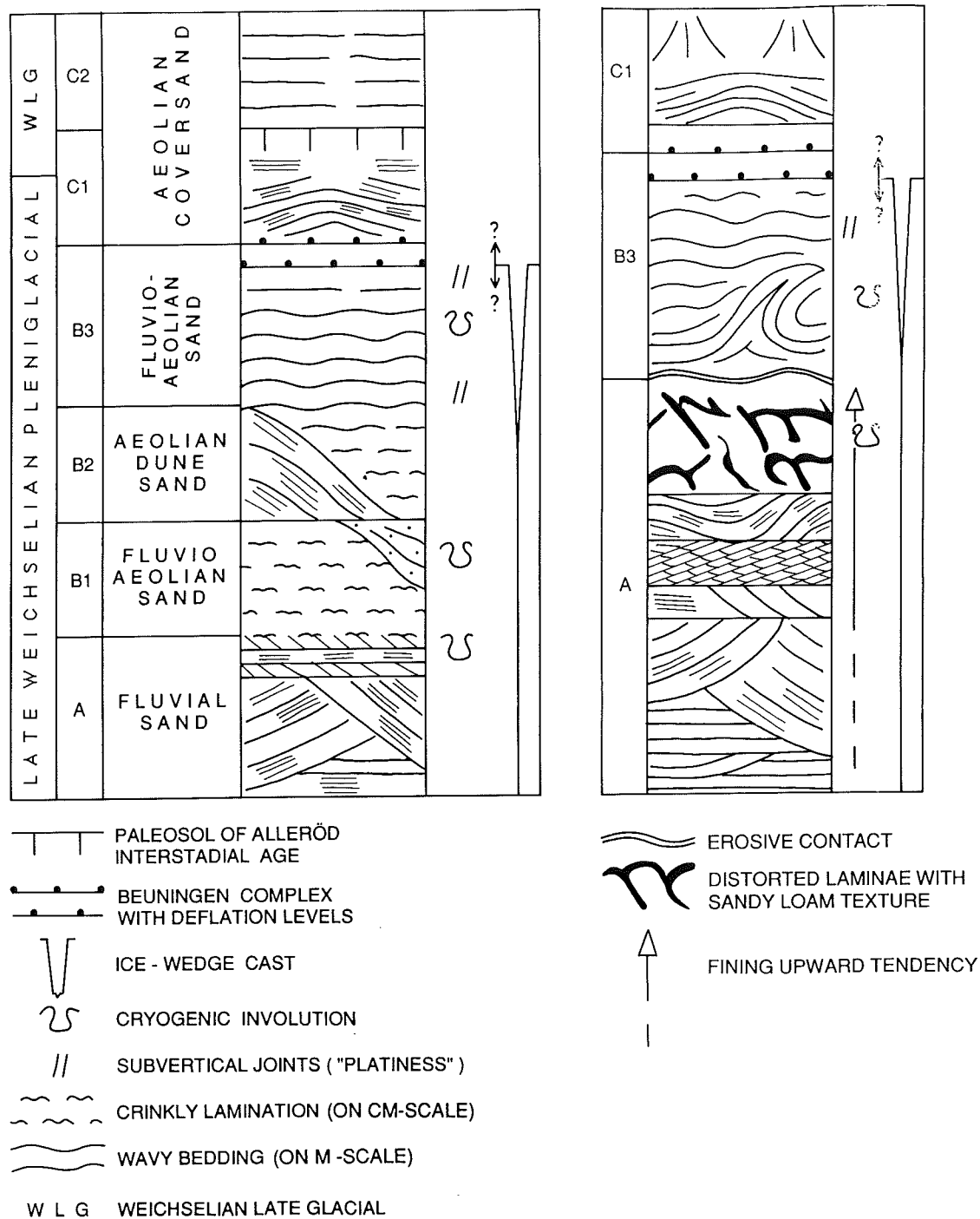


Fig. 3 Schematic sedimentary succession in sand pit Grubbenvorst in August 1988 (left side) and Novembre 1990 (right side) (taken from Schwan & Vandenberghe, 1991).

From borehole-observations in the vicinity of the exposure, it is known that a succession of three such sequences is present in the deeper subsoil and extends to a depth of approximately 7.5 m below the floor of the sandpit.

Occasionally, within unit A, channels are found whose infilling, rather than being concordant, is lateral; the attitude of the foresets suggests that they were formed by a streamflow which was roughly perpendicular to the axis of the host channel.

Besides channel sediments, unit A contains horizontally bedded units with large-scale tabular cross-bedding, which are attributed to deposition by straight-crested bars on the floodplain (fig. 4). The dip of the foresets indicates a stream direction towards the northwest or north. The silt drapings within the bar foresets reflect periods of low current velocities or standing water conditions following high current velocities during flood.

The unit-A deposits are believed to have formed by an aggrading braidplain. The large-scale channel fills are suggestive of episodic peak discharges with erosive streampower, - seasonal or otherwise, - followed by a phase of infilling when the flow strength had diminished. The wide lateral spread of the channel occurrences is most likely to be a result of shifting behaviour of the ancient Maas. Because of the threefold succession alluded to above, some kind of cyclicity must have been involved in the buildup of the braidplain.

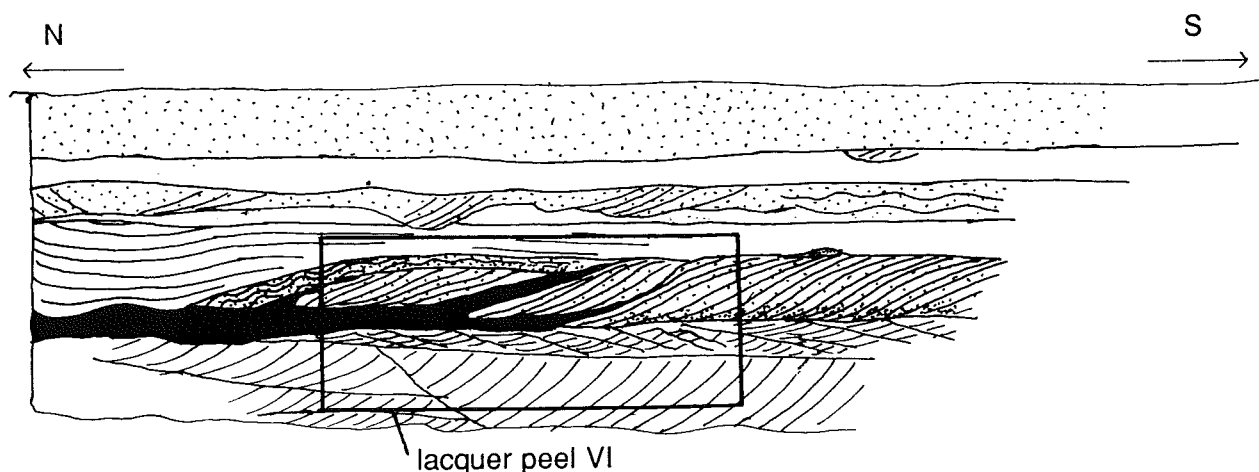


Fig. 4 Silt drapings on the foresets of large-scale transverse bars at the base of fluvial unit A. These lateral changes illustrate the recurrent alternation of sand deposition during peak discharges and silt deposition during standing water conditions on the braided floodplain.

Units B1 to B3

Whereas units B1 and B3 correspond to the fluvio-aeolian type, unit B2 is a deposit of small dunes having a purely aeolian origin (Fig. 3). As far as known, the occurrence of unit B2 is restricted to a single stretch less than 10 m in length which,

by now, has fallen prey to excavation. There, the unit had a thickness of 1.5 m only and, by attitude-readings on the dune-foresets, deposition from a west-northwestern direction could be deduced. Wherever unit B2 is absent, the upward transition from unit B1 to B3 is gradational.

Unit B1, the lower fluvio-aeolian unit, has an overall granular composition of fine sand with a few thin intercalations of either silty or coarser-than-average sandy texture; layers of the latter type are lense-shaped with a concave lower boundary and a flat upper one. Whereas finely-laminated parallel bedding is the prevailing stratification type, low-angle cross-lamination, ripple-foreset cross-lamination and small scour-fill structures are subordinate characteristics that occur regularly. Normal grading from (fine) sand to a more silty texture is frequently found in one to two cm thick sets. Almost certainly as a result of cryogenic stress, all these features have a generally crinkled appearance when viewed on a cm-scale. Moreover, one to three cm thick micro-loadcasted levels add to the impression of small-scale distortion.

Channel fills laterally merging with the parallel-laminated sands occur here and there in unit B1; their scarcity testifies to the dwindling importance of fluvial activity.

Unit B3, the upper fluvio-aeolian unit, has a make-up quite similar to that of B1, except for the absence altogether of fluvial channel-fills. In unit B3, laminae are organised in rather thick sets separated by indistinct bedding planes. Owing to this, the waviness of the beds on a m-scale stands out clearly. Like the crinkliness discussed before, this feature must be attributed to cryogenic stress with the difference between the two being merely a matter of size.

Units B1 and B3 are thought to be a result of fluvial aeolian interaction as, in the subject sediment, planebedded layers of generally accepted aeolian origin (e.g. Schwan, 1988) occur side by side with features that strongly suggest aquatic conditions with or without current flow. In elaboration of the above it is assumed that (i). Buildup of the fluvio-aeolian sediment takes place on an abandoned braidplain which is flooded, now and then, by the river and (ii). Deposition of the blown-sand component proceeds, mainly, in sheetlike fashion, so that the former braidplain will retain its overall horizontality whilst it aggrades. When these requirements are met, the following processes may become operant:

1. Aeolian deposition of a thin sheet of sand on a dry or damp land surface,
2. Sand grains driven by wind or fines falling out from the air which come to rest in a temporary and shallow pool of still water,
3. Reworking of previously deposited material by high-energy sheet flows during a flooding event.

Clay-rich mud drapes that often characterise sandy overbanks subject to flooding (e.g. Langford, 1989) are rare in the sediment under consideration. Possibly, this is due to a primary lack in very fine particles in the suspension load of the river.

Locally, units B1 and B3 are separated by an aeolian dune facies. As already told, the dunes are low and insignificant in areal extent. This data, therefore, does not contradict the assumption of aeolian deposition mainly in the form of low-angle sand sheets.

Unit C

Unit B3 is capped by, successively, the Beuningen Complex and a packet of aeolian coversands (unit C) (Figs. 2 and 3). Whereas the former type results, mainly, from deflation, the second one is a characteristic product of aeolian deposition. Unit C1 is characterized by horizontal lamination and low-angle cross-bedding and is assumed to be deposited in a sand sheet of low dune environment. Unit C2 is separated from Unit C1 by a reddish brown sandy silt, which on stratigraphical grounds may be equivalent with the Usselo soil of Allerød age. The soil represents a phase of local surface stability. Locally, the transition between the sand units C1 and C2 is not a single bed of sandy silt, but rather an interval with alternating bedding of fine sand and silt. Rather than a period of complete absence of aeolian deposition and associated soil formation, this alternating bedding suggests a period of decreased aeolian deposition.

Cryogenic deformation structures

In the Grubbenvorst exposure, four different types of cryogenic deformation features are observable. These are:

- * Casts of epigenetic and syngenetic ice wedges (see Fig. 5)
- * Cryoturbation structures (= periglacial involutions) (see Fig. 2)
- * Warping and crinkling of originally evenly bedded strata.

In the foregoing, the corresponding deformations were termed "waviness" and "crinkliness" respectively. Though the latter category is attributed to cryogenic stress in the first place, it cannot be excluded that microloading of water-saturated sediment is another agent capable of producing crinkly structures.

- * Sets of subvertical joints developed in an en-echelon pattern (see Fig. 6).

In literature, the sets of subvertical joints are sometimes referred to as "platiness". Mol et al. (1993) attributed the development of these vertical platy structures to intense cooling and cracking of the surface in periglacial environments. Stratigraphically, the vertical platy structures are most common in Weichselian Upper Pleniglacial beds. They generally occur in association with ice-wedge cast and therefore are assumed to have been generated under the very severe cold conditions of the last glacial maximum.

With respect to the above it is noted that:

- 1). The degree of development of anyone of the cryogenic features is laterally variable. This is to say that, within one and the same unit, a phenomenon like cryoturbation may be well expressed in one place and lack in another one. Granular composition and height above the groundwater table are two often quoted parameters to account for this variability.
- 2). Periglacial deformation structures occur at several different levels in the exposure. Their morphology and stratigraphic position is as follows:
 - (i). Three levels of involutions with amplitudes of respectively 40, 50 and 80 cm in the

top of unit A,

(ii). Strongly cryoturbated channel fills with silty texture in unit B1,

(iii). Waviness locally changing into involuted structures in unit B3,

(iv). Ice-wedge casts departing from both the top of unit A as well as from unit B; in the lower part of the exposure, some of the casts have been truncated by fluvial channels.

3). The strata directly overlying the youngest ice-wedge casts are entirely free of large-scale distortion, except for their wavy and crinkly appearance. Normally, melting of the ice wedges is accompanied by involution of the waterlogged active layer. Absence of the resultant structures suggests either removal by erosion of the affected levels or inactivity of the periglacial-loadcasting process when the permafrost degraded. The second alternation is thought to apply to the Grubbenvorst site.

4). The strong development of the cryogenic features in units A and B of the Grubbenvorst section is in full agreement with the chronostratigraphic age attributed to them (see section Stratigraphy). Following various sources (e.g. Vandenberghe, 1985) the early part of the Late Weichselian Pleniglacial in northwestern Europe was a period of continuous permafrost corresponding to the second cold maximum of the Last Ice Age.

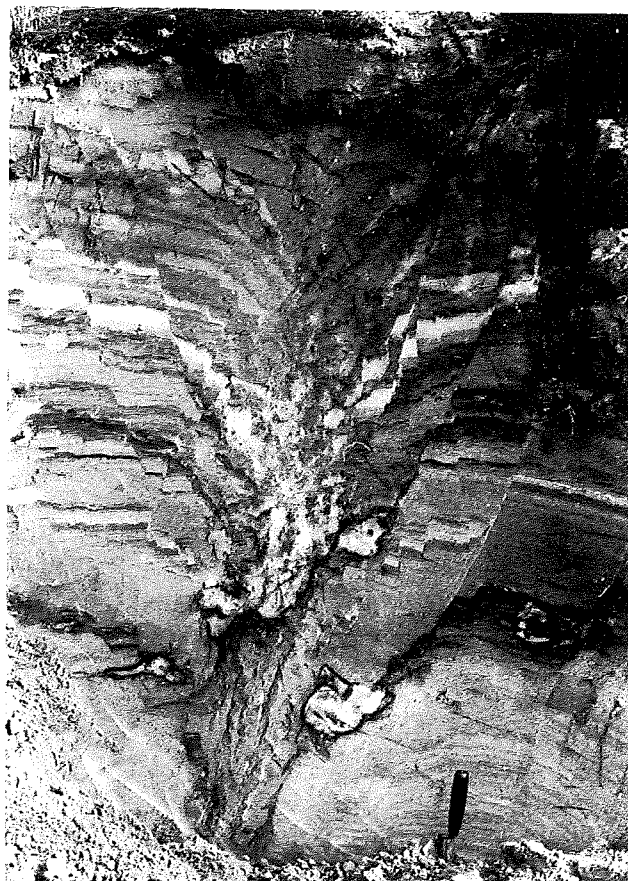


Fig. 5

Ice-wedge cast in fluvial unit A at Grubbenvorst demonstrating the presence of continuous permafrost during the Weichselian Late Pleniglacial (trowel is 25 cm) (from Mol et al., 1993).



Fig. 6 Vertical platy structures in the fluvio-aeolian unit B at Grubbenvorst showing different directions of the microjoints (knife is c. 25 cm) (from Mol et al., 1993).

Palaeo-environment

In the period from 26 ky BP to approximately 18 ky BP, conditions of continuous permafrost prevailed at the exposure site and its wide surroundings. Concurrently, the depositional regime changed from a fluvial one to a fluvio-aeolian one, possibly caused by stronger aridity towards the end of this period.

In the timespan from 18 ky to c. 14 ky BP, degradation of the permafrost was followed by a hyperarid phase with predominance of wind activity; in that interval, the desert pavement of the Beuningen Complex came into being. A subsequent amelioration of climate resulted in the deposition of windblown sand (Kolstrup, 1980). Possibly during the Allerød Interstadial of the Weichselian Late Glacial, the aeolian accumulation process slowed down or was interrupted by pedogenesis.

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EXCURSION STOP 2: BOSSCHERHEIDE

Introduction

Bosscherheide is located on the east bank of the Meuse in an extensive dune area, characterized by its general parabolic outline (Fig. 1). As a result of intensive exploitation of the area for its gravels and sands numerous profiles became available in which fluvial and aeolian environments, dated to the Weichselian Late Glacial, could be studied (see Fig. 2). Well-developed Younger Dryas cryoturbation structures have been found at the transition of both environments (Fig. 6 and 7). The organic infill of an abandoned gully and its lateral extension over the adjacent terrace (see Fig. 2) has been sampled for pollenanalyses and dating purposes (Figs. 3 and 4). The cryoturbation structures have been studied macroscopically (Fig. 7).

It is quite common in the Dutch and German lowland region that, buried under parabolic dunes of Younger Dryas age, a palaeosol of Allerød age, the so-called Usselo soil (Van der Hammen, 1951) occurs. This is also the case at excursion site Bosscherheide where, at the base of the dune sediments, Late Glacial organic sediments and peats are present.

Sandwiched between the base of the dune and the organic sediments the lithological sequence at Bosscherheide reveals the results of a series of processes during the Younger Dryas, that previously had been observed at a site called Notsel (Vandenberghé et al., 1987 and Bohncke et al., 1987). In Notsel, however, dating was difficult due to the absence of organic material in the Younger Dryas interval. Two important features were described at Notsel and placed in the early Younger Dryas being, 1) a reactivation of the fluvial regime visualized by the deposition of fluvial loams and sands and 2) the local establishment of deep seasonal frost or discontinuous permafrost as witnessed by the occurrences of a transitional form between an ice-wedge cast and a frost fissure (Fig. 5).

Dating of the terrace at Bosscherheide

In the area between Arcen and Bergen the terraces are partly of Rhine partly of Meuse origin (Zonneveld, 1956). At Bosscherheide the substratum is formed by the Well Sands, which have a Rhine heavy mineral assemblage. According to Zonneveld (1956, 1958) these gravels and coarse sands were deposited in the period in which the expanding Saalian inland ice forced the Rhine to divert its course in westerly directions; the depressions between the terraces of Walbeck and Twisteden would have functioned as the southern most escape route of the Rhine waters. During the Weichselian Middle and Late Pleniglacial these Rhine deposits have been eroded and reworked by a braided system to form the Kreftenheye Formation (Pons, 1957; Teunissen, 1983). The Well Sands of Saale age are still present in the eastern part of the exposure, while the Kreftenheye Formation is found in a somewhat lower position in the western part of the pit.

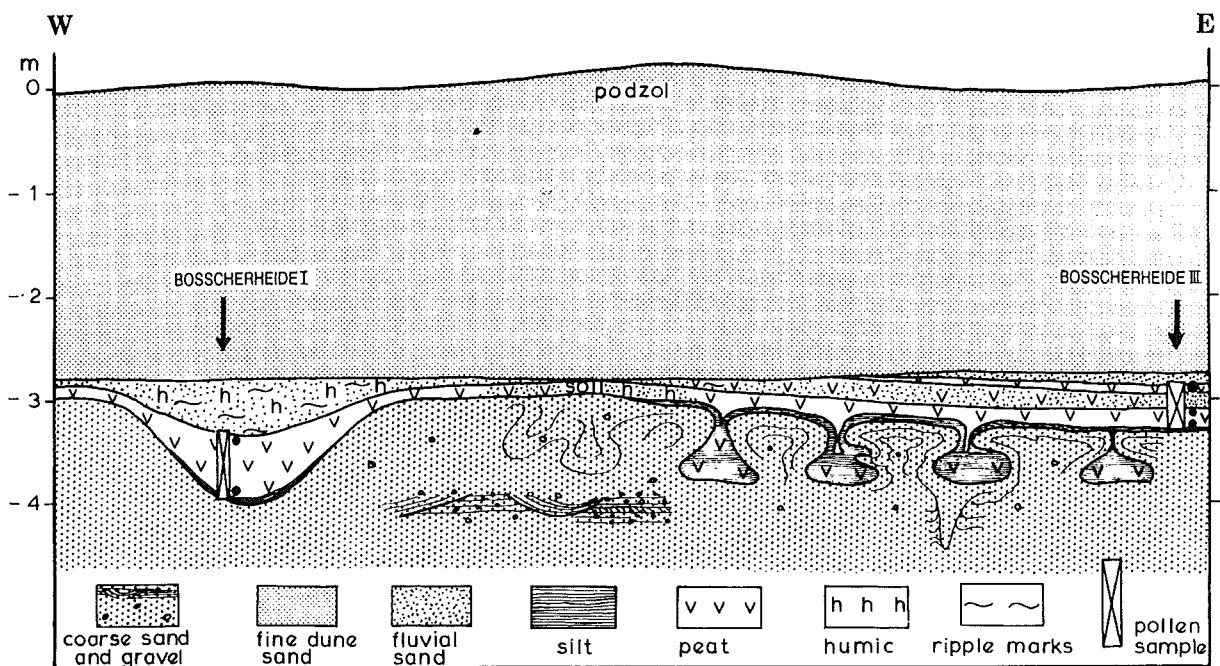


Fig. 2 Schematic profile of exposure Bosscherheide showing the location of the pollen samples. The black dots refer to radiocarbon-dated levels (taken from Bohncke et al., 1993).

Bosscherheide I from top to bottom:

- $10,940 \pm 60$ BP
- $11,300 \pm 60$ BP
- $12,100 \pm 70$ BP
- $12,110 \pm 70$ BP

Bosscherheide III from top to bottom:

- $10,500 \pm 60$ BP
- $10,880 \pm 50$ BP
- $11,500 \pm 50$ BP

A dominant feature of this Late Pleniglacial fluvial unit is the quick alternation of gullies. Many gullies contain horizontal, sharply bounded sets of sand and gravel. These relatively thin sets often show an internal oblique lamination and may be interpreted as channel bars. Other gullies are filled completely with obliquely laminated sands, caused by lateral accretion. The described sedimentary structures point to a high accumulation rate in a multi-channel river system.

This fluvial unit is capped by a sandy grey loam at the time the fluvial system merged into an incising single-bedded channel, transforming in this way the earlier fluvial deposits into a terrace. This terrace forms part of the Meuse lower terrace (see also Teunissen, 1983). In the abandoned gullies organic sediments started to accumulate. Virtually no time seems to have elapsed between the abandonment of the gully and the first accumulation of organic matter, since aquatic species are abundant in the bottom most samples of the infill (see Fig. 3, the pollen record).

The pollenrecord Bosscherheide I and BH III; chronology and paleoenvironment of the Late Glacial sediments (Figs. 3 and 4)

Biostratigraphically registration in the abandoned gully of Bosscherheide starts with the Bølling (Fig. 3: zone 1b and 1c) as demonstrated by the high *Betula* values in these zones. 7,5 km southeast of Bosscherheide, Teunissen (1983) obtained the oldest radiocarbon date of $12,760 \pm 150$ BP (GrN-4478) for the transition between the underlying thin sandy loam and the organic channel infill. A similar date ($12,600 \pm 60$ BP) has been obtained from the organic infill of an abandoned braided channel at Notsel (Bohncke et al., 1987).

The relatively high *Betula* values show a sudden drop at the transition to the Older Dryas at the base of zone 2 (Fig. 3). The top of the Bølling has been dated here at $12,100 \pm 70$ BP. The Older Dryas is characterized by a herbaceous rich pollen assemblage in which *Artemisia* is frequently encountered. Subsequently, *Betula* gradually restores (Fig. 3: zone 2 *Betula*-phase) after which *Pinus* becomes the dominant species in the vegetation cover (zone 2 *Pinus*-phase). Zone 2 reflects the *Betula*- and the *Pinus*-phase of the Allerød.

The lateral extension of the peat over the adjacent terrace has been studied in a site similar to Bosscherheide III (see Fig. 2) and it reveals the *Pinus*-phase of the Allerød (Fig. 4: zone 2) followed by a herbaceous rich vegetation (Fig. 4: zone 3), where the lithology changes into a fluvial loam. More distal from the alluvial plain of the Maas this Allerød peat changes into a sandy, charcoal rich soil, but the overlying fluvial loam of Younger Dryas age remains a consistent feature.

The top of the Allerød peat has been dated in the gully and on the adjacent terrace providing dates of $10,940 \pm 60$ BP and $10,880 \pm 50$ BP respectively. At Notsel the termination of the Allerød was dated to $10,970 \pm 50$ BP and south of Bosscherheide, Teunissen (1983) obtained a date of $10,870 \pm 100$ BP. These radiocarbon dates unambiguously put the overlying fluvial loam in the early Younger Dryas period.

At the transition to zone 3 (Figs. 3 and 4) the *Pinus* forest suddenly drowns as appears from the return of aquatic species and the deposition of fluvial loams over the interstadial peat. The reactivation of the fluvial system is clearly demonstrated by the occurrence of *Classopollis*, an element of the Tertiary brown coal layers, in the pollen record. *Azolla tegeliensis* was encountered in the fluvial loams that cap the interstadial Bølling/Allerød peat at Notsel (Bohncke et al., 1987) emphasizing the regional occurrence of this process of reactivation of the fluvial regime during the early Younger Dryas. The lateral extension of the Younger Dryas loam reflects large-scale floodings and deposition of suspension load in large shallow lakes (Pons & Schelling, 1951).

Fig. 3 Pollen diagram and macro remains diagram (selection of curves only) of Bosscherheide I taken from the abandoned channel.

Fig. 4 Pollen diagram of Allerød soil and Younger Dryas loam and peat adjacent to the abandoned channel in a position similar to Bosscherheide III (see Fig. 2).

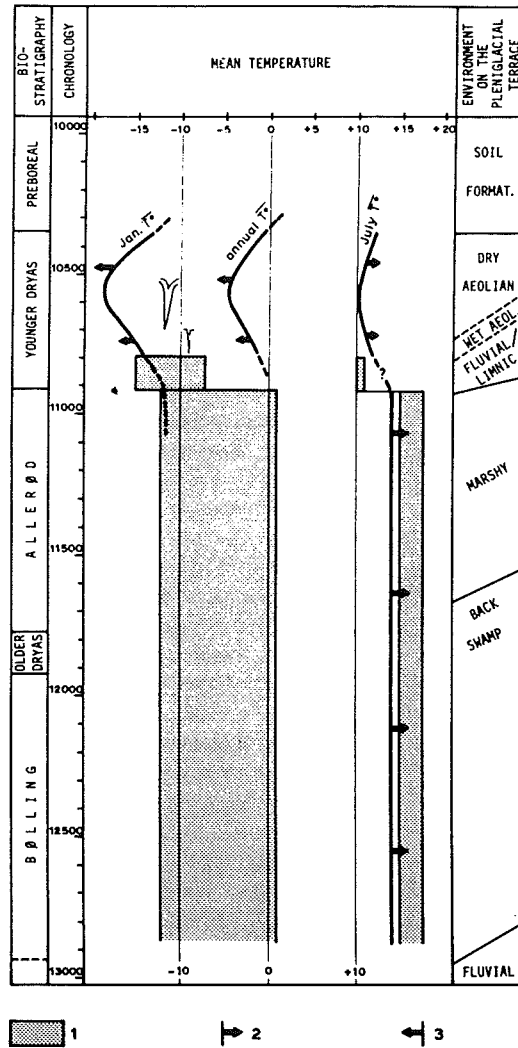


Fig. 5 Tentative reconstruction of temperature fluctuations during the Late Glacial based on sequence Notsel (Bohncke et al., 1987). The arrows correspond to the levels with specific temperature indicator species within the peat section at Notsel, based on Coleoptera (1), palaeobotanical evidence (2) and periglacial phenomena (3).

The profiles at Bosscherheide show that, subsequent to the deposition of organic matter during the Bølling and Allerød and preceding the deposition of the overlying fluvial loam, a period with deep seasonal frost or incipient permafrost established (Figs. 6 and 7). According to the ^{14}C -dates provided by profile Bosscherheide III, where the top of the Allerød peat had been dated, this local permafrost established directly following 10,880 BP (see below). One may assume that the local permafrost could have installed already in due course of the Allerød, but this is strongly contradicted by the temperature curve from the peat sequence at Notsel (see Fig. 5).

With the deposition of fluvial loams on top of the Allerød layer the subsoil thawed, leaving nicely developed loading structures (see Figs. 6 and 7). The loam, although partly filling the morphology that established following the permafrost degradation, is horizontally laminated and hence postdates the phase with permafrost. Another argument for the fact that permafrost degraded during or preceding the deposition of the Younger Dryas loam is the absence of segregation ice in the loam, whereas this feature is present in the underlying organic deposit and Late Pleniglacial fluvial loam as appears from the study of thin sections of the deposits (see below: Periglacial features).

More distal these Younger Dryas fluvial deposits are intercalated by thin moss layers and clayey gyttjas and peat. This sediment series represents the end of the fluvial activity at this location. On top of this sequence an aeolian layer of varying thickness and consisting of horizontally bedded and low-angle cross-bedded sands has been deposited. The start of the aeolian activity at this location could be dated by the moss layer directly underlying the aeolian deposits (Figs. 4 and 6). A date of $10,500 \pm 60$ BP has been obtained, and places the aeolian activity at this location in the late Younger Dryas.

Periglacial features

The upper part of the terrace sands and gravels and the lower part of the overlying fine-grained, organic layer are deformed by a regular pattern of well-pronounced involutions (Figs. 6 and 7). The mean amplitude is ca. 60 cm. The updoming parts have a diapir-like form and the sinking parts reach a remarkably constant depth (Fig. 6). Comparison with similar involution structures allows the interpretation as load casts (Vandenberghe & Van de Broek, 1982). The horizontal boundary between the flat-bottomed involutions and the underlying, actually very permeable sediments may be interpreted as a former permafrost table. Although the amplitude of the involutions is relatively small in comparison to cryoturbations from other sites in the same region (Vandenberghe & Krook, 1981 and 1985), the very regular development of the involutions testifies of their single development and excludes a multiple formation, for instance as a recurrent process in yearly thawing soils (French, 1986).

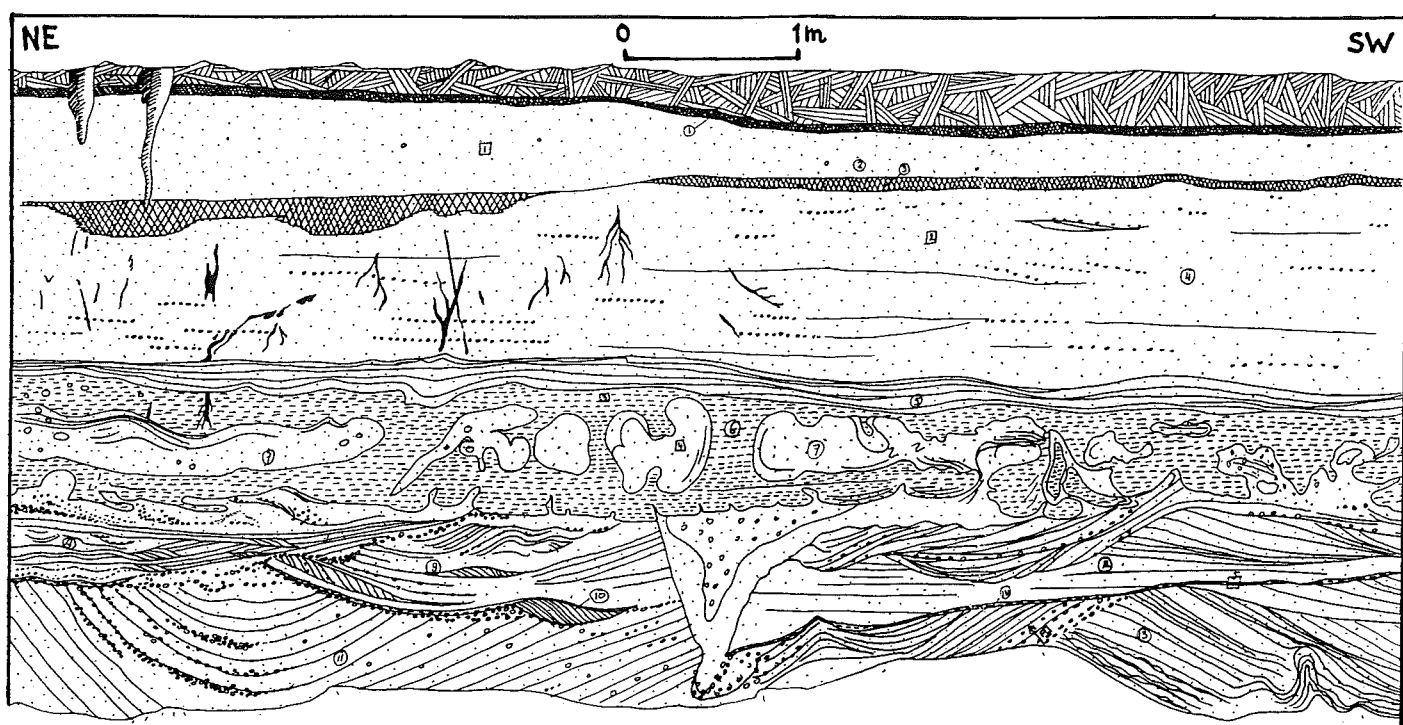


Fig. 6 Drawing of a profile face at Bosscherheide showing a) Late Pleniglacial to early Late Glacial braided sediments, b) Younger Dryas involutions and initial ice-wedge cast and c) late Younger Dryas dune sands.

The generation of the described involutions on top of a degrading frozen subsoil may further be argued by the wedge forms just below the involution zone (Fig. 6). They reach a depth of 45 cm and a maximum width of 15 cm. Very distinct upturning has been observed, but downturning occurs as well. The infilling with completely homogenized sand sharply contrasts with the surrounding distinctly bedded sediments. The limited depth of these wedges (c. 1 m from the former surface) and the absence of subsidence features (faults and slump structures) questions their interpretation as ice-wedge casts. The absence of vertical lamination and the only weakly upturned edges reject an interpretation as sand wedge casts. A similar wedge form, but the upper part not considerably disturbed by cryoturbation, has been described by Vandenberghe et al. (1987) at Notsel in the Mark valley c. 100 km to the west. It could be interpreted either as an intensely developed frost fissure, formed during numerous successive winters in a deep (min. 1 m) seasonal frozen ground or as an initial form of ice-wedge cast. Given the considerations about the overlying cryoturbation zone, the latter interpretation now seems more probable. It is concluded that in this region local permafrost was present for a very short time.

Shortly after the intense cryoturbation, the soil underwent again an involution process, but the resulting deformations - although regularly developed - have a very

small amplitude (max. 30 cm) and a large wavelength (c. 120 cm). Small slump structures occur on the steeper parts of the undulations. The described forms resemble "earth hummocks" (f.i. Tarnocai & Zoltai, 1978) and "thufurs" (f.i. Schunke, 1977). Besides, the undulations at Bosscherheide consist of a peat and a peaty loam layer at their top as do thufurs and earth hummocks. It is, however, at present not clear if the same processes are involved.

Thanks to radiocarbon analyses of surmounting and underlying peat layers the periglacial phenomena can be dated in detail. The rate of peat growth of the main peat layer could be calculated from the radiocarbon dates of $10,880 \pm 50$ BP and $11,500 \pm 50$ BP in Bosscherheide III (see Figs. 2 and 7). It follows that the peat formation ended at c. $10,750 \pm 50$ BP. The intense cryoturbation took place just after this peat formation. An overlying peat layer, not affected by cryoturbation, is dated at $10,500 \pm 60$ BP. Assuming the same growth rate the latter peat formation started at c. $10,550 \pm 50$ BP. Thus the degradation of the frozen layer, manifested by the intense cryoturbation, took place between 10,550 and 10,750 BP. Also the later formation of weaker undulations took place before $10,550 \pm 50$ BP. The age of the (local) permafrost has to be situated shortly before that cryoturbation phase. It has to be stressed that all these periglacial phenomena may be formed in a short time. The described wedge form may, according to its maximum width of 15 cm and an assumed mean growth rate of 0,5 à 1 mm/yr be formed in 150 to 300 years. Involutions on top of a frozen soil are due to loading which is an instantaneous process during the melting of the frozen subsoil (Vandenberghe, 1983).

Finally, the degradation of the (local) frozen subsoil is not necessarily the expression of a climatic amelioration. Especially in valleys the talik underneath the river may move laterally causing the melting of the neighbouring soil (Bryant, 1983). This seems likely at Bosscherheide since the Maas river plain was nearby. According to the micromorphological analyses, however, permafrost did not return afterwards. In the Mark valley at Notsel, the above mentioned wedge has developed and degraded also shortly before the dune formation which started probably around 10,450 BP. As a conclusion it may be derived that the coldest phase of the Younger Dryas has to be situated between c. 10,800 and 10,500 BP.

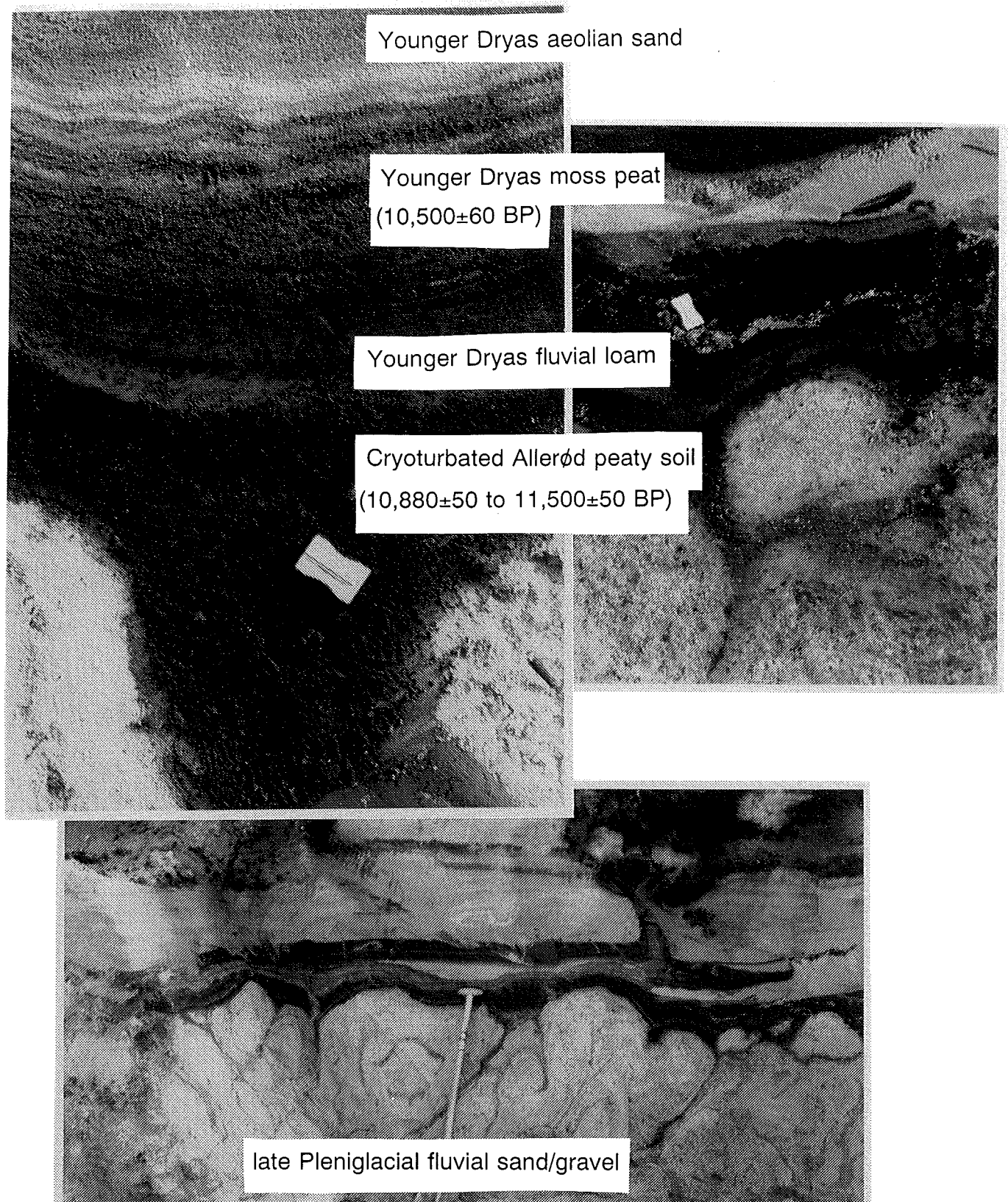


Fig. 7 Younger Dryas involutions at Bosscherheide dated between c. 10,880 and 10,500 BP

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EXCURSION STOP 3: LATE ALLERØD MEANDER SCAR AT MEERLO

The meander scar at Meerlo is situated at the same Late Glacial terrace as the one at Beugen (Excursion stop 5). The scar is up to 175 m wide and the fine-grained infill is up to 5 m thick. The meander scar at Meerlo differs from the one at Beugen in that it has been abandoned at the start of the Younger Dryas, possibly by the process of meander chute cut-off. The Younger Dryas floodplain is located less than 1 km north of the meander scar (see Fig. 1 and Excursion stop Kasteelweg-Ooijen, Fig. 3).

The meander has been abandoned due to the rapid changes in the fluvial regime at the Allerød-Younger Dryas transition. Peak discharges, possibly due to snow melt in late spring, led to a strong modification of the river channels and the establishment of a new balance between channel morphology and climate (see Excursion stop 4: Kasteelweg-Ooijen). A more straight and wide (multi-) channel floodplain was required for the larger peak discharges that characterized the climate at the time of the chute cut-off.

During the higher Younger Dryas peak discharges the abandoned Allerød channel of Meerlo was still flooded and here deposition of fines from suspension load took place (see lithology of the pollen record in Fig. 2). The fine-sandy clay at the base of the meander infill reflects this episode of peak discharges straight after the cut-off. The scar infill reveals a fining-upward sequence. The sandy clay at the base gradually changes into humic clay and gyttja. The clay is calcareous and locally contains ostracods, fresh water snails and burrowing structures.

The pollenassemblage (Fig. 2) shows relatively high values for the Non Arboreal taxa, among which the grasses and the sedges are dominant. Furthermore, the relatively high values for the *Artemisia* and the continuous presence of *Empetrum* are indicative for the **Younger Dryas** chronozone (Fig. 2: 350-520 cm). The presence of *Classopollis*, a reworked taxa from the Lower Cretaceous beds in the upper course of the Maas, characterizes the clastic part of the infill during which the straight and wide (multi-) channel floodplain of the Younger Dryas was active.

Resuming, the Allerød large meandering system was abruptly abandoned as a result of swift morphodynamic adaptation of the fluvial regime, which in turn was forced by a dramatic climate cooling. This change in climate involved an increase in precipitation, predominantly in the form of snow.

At the Younger Dryas-Preboreal transition a change in the lithology from humic silty clay to gyttja is present. *Classopollis*, derived from Cretaceous beds upstream in the Maas catchment, and other reworked species (e.g. *Corylus*, *Quercus*) disappear at this level. These changes in lithology and pollen assemblage indicate a decrease in the flooding frequency of this site by the Maas river.

The reasons for these changes at the Younger Dryas-Preboreal transition are a series of processes involving climate improvement, the related change in fluvial regime from braided back to meandering and incision of the river. The nearby western branch of the Younger Dryas braided system became abandoned, as has been established at Excursion stop 4 (Kasteelweg) (see Fig. 1 for locations). The eastern branch of the Younger Dryas system near Ooijen (Fig. 1) became the principal river course and developed into the early Holocene floodplain (see Excursion stop 4: Ooijen). Because of these hydrological changes the Meerlo meander scar became situated more than 3 km from the Preboreal floodplain, in contrast to only 1 km during the Younger Dryas. The distal position of the meander scare in combination with the early Holocene incision evidently led to the changes in the sediment and pollen record at this level (Fig. 2).

The **Preboreal** (Fig. 2: 268-350 cm) zone starts with a distinct lacustrine environment with high values for the aquatic taxa. The pollen record shows a nice succession with *Juniperus* and *Betula nana*, followed by tree birch and pine. The increase in *Pinus* in the Maas area has been dated to $9,750 \pm 50$ BP. The local hydrosere succession at this time reached the semi-terrestrial phase, leading to the formation of peat. At the top of the pollen record (Fig. 2: 240-268 cm) the increase in autochthonous *Corylus* pollen determines the transition to the **Boreal** (c. 9100 BP).

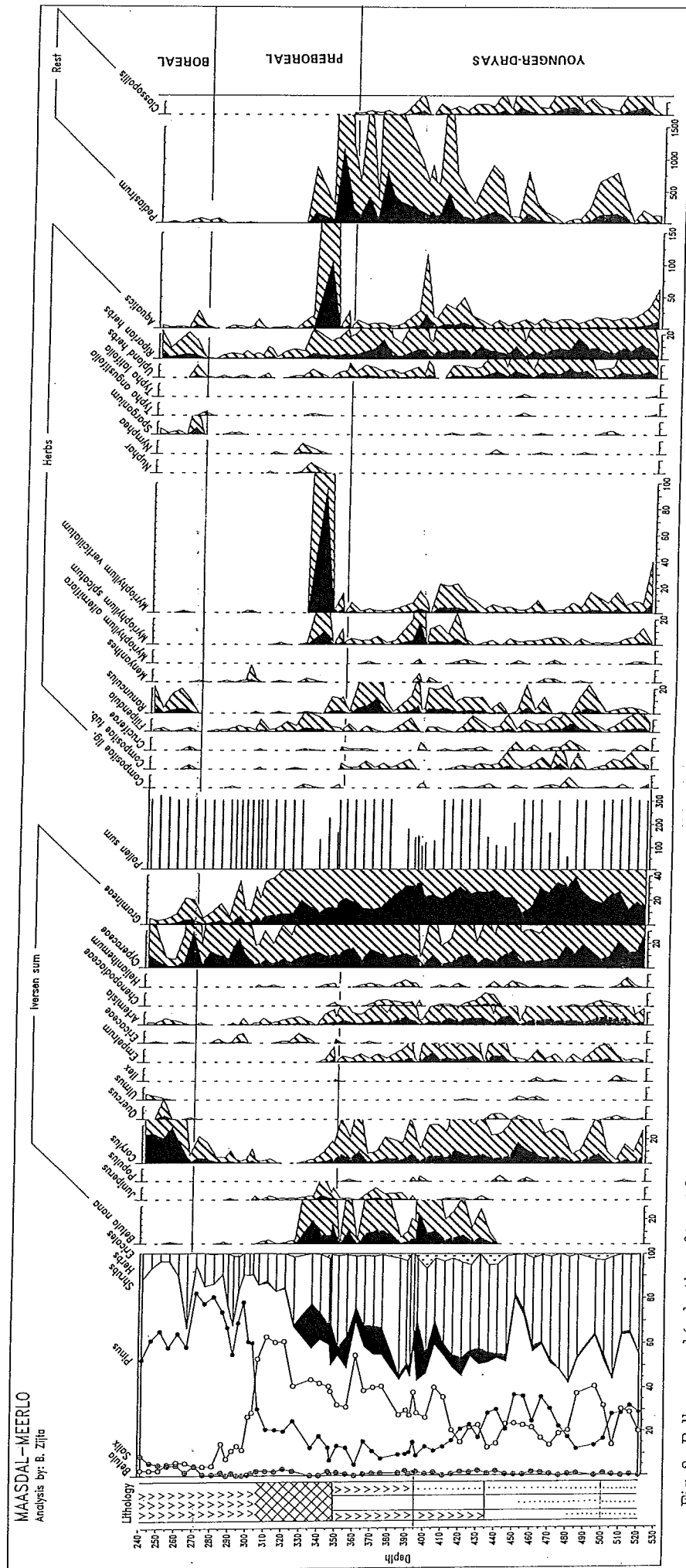


Fig. 2 Pollen record (selection of taxa) from the infill of the meander scar at Meerlo.

EXCURSION STOP 4: YOUNGER DRYAS AND HOLOCENE FLOODPLAIN AT KASTEELWEG-OOIJEN

Introduction

At this excursion point a 1,5 km walk will be made, in which we cross the Younger Dryas and Holocene floodplain. Kasteelweg-Ooijen is located on the western bank of the Maas. The altitude is between 14 (channels) and 20 meters (river dunes) above sea level (Fig. 1).

In this region the Younger Dryas cooling of the climate has been registered in the following manners (Fig. 2) (Kasse, in press):

1. Change in channel pattern. The Younger Dryas floodplain is generally 1 km wide and characterized by a braided floodplain with straight to low sinuosity channels (Fig. 3). The change in channel morphology from high sinuous to braided probably occurred during the start of the Younger Dryas, because the temperature decrease led to lower evapotranspiration and higher peak discharges. The morphological change was accompanied by erosion of the high-sinuosity meander terrace (see Fig. 4).

2. Chute cut-off of meanders. The rather straight Younger Dryas floodplain was formed by several chute cut-offs of high-sinuosity Allerød meanders (Fig. 4). This is concluded for instance from a nearby meander (Meerlo, 2.5 km west of this excursion point) which was abandoned and filled with sandy clay during the Younger Dryas.

3. Flooding of older terraces. Because of the higher peak discharges the Maas was able to inundate areas that were previously above the flood limit, for instance higher lying Late Glacial or late Pleniglacial terraces. At Bosscherheide (Excursion stop 2) a fluvial loam was deposited over the Allerød soil/peat and the Younger Dryas involutions.

4. Lithological change in abandoned channels. Because of the higher peak discharges, the Maas was able to inundate not only older terraces but also abandoned meander channels of the Allerød period. At Beugen (Excursion stop 5) a gyttja of Allerød age at the base of the infill is erosively overlain by a gray fine sandy clay. This lithological break correlates pollenanalytically with the start of the Younger Dryas.

5. Widespread aeolian deposition. On the east bank of the Maas valley an up to 4 km wide sand-sheet and dune belt is present, overlying older Late Glacial or late Pleniglacial deposits. The parabole dune morphology indicates a westsouthwesterly wind during their formation in the late Younger Dryas. At Bosscherheide (Excursion stop 2) the aeolian sediments overlie a peat layer dated at $10,500 \pm 60$ BP. The reason for this large-scale aeolian deposition is related with the change in channel morphology at the Allerød-Younger Dryas boundary. The single-channel meandering system was at that moment replaced by a much broader multi-channel braided system. During low discharge, sediment could be deflated from this wide south-north oriented Maas floodplain by the westerly wind.

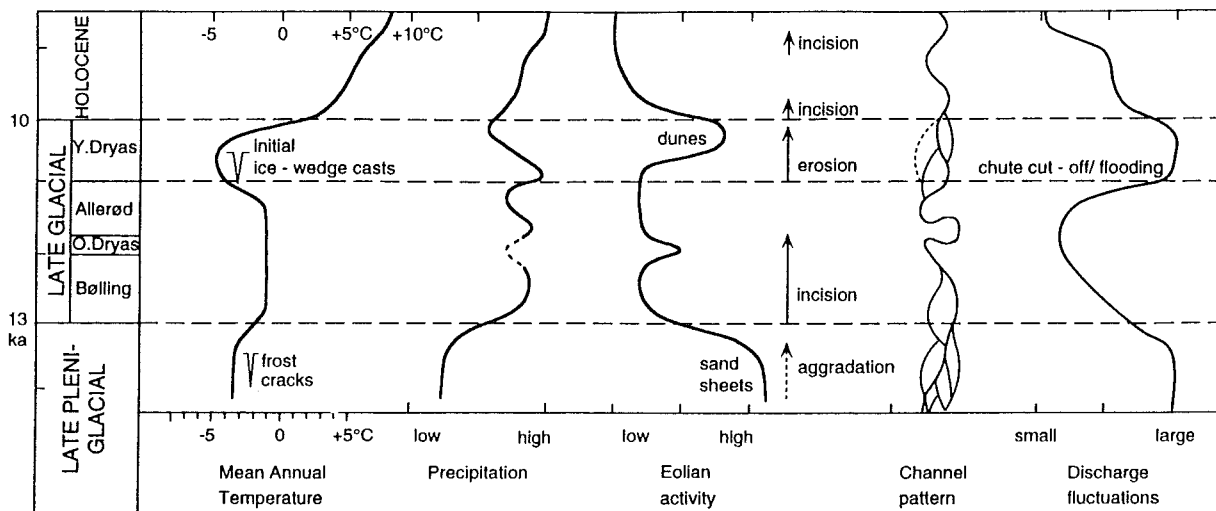


Fig. 2 Summary of the Late Glacial climatic and environmental changes in the Maas valley (from Kasse, in press).

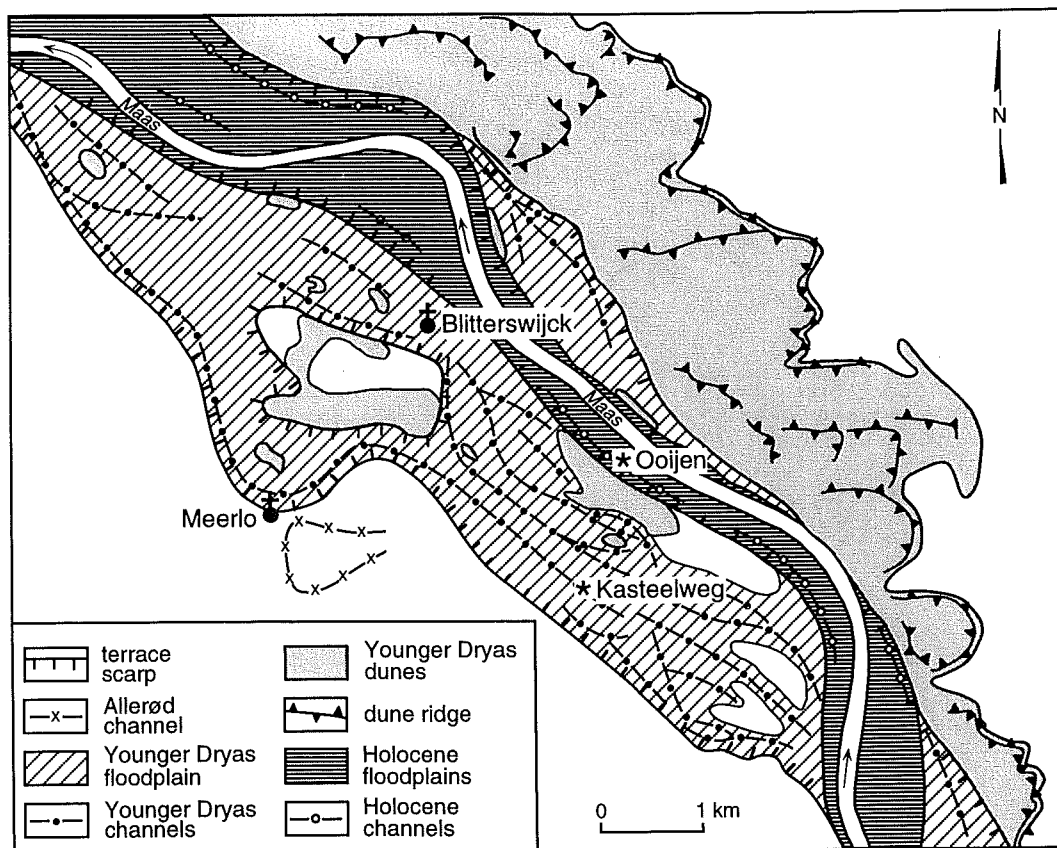


Fig. 3 Morphological map of the Younger Dryas braided floodplain and Younger Dryas dune field at Blitterswijk (from Kasse, in press).

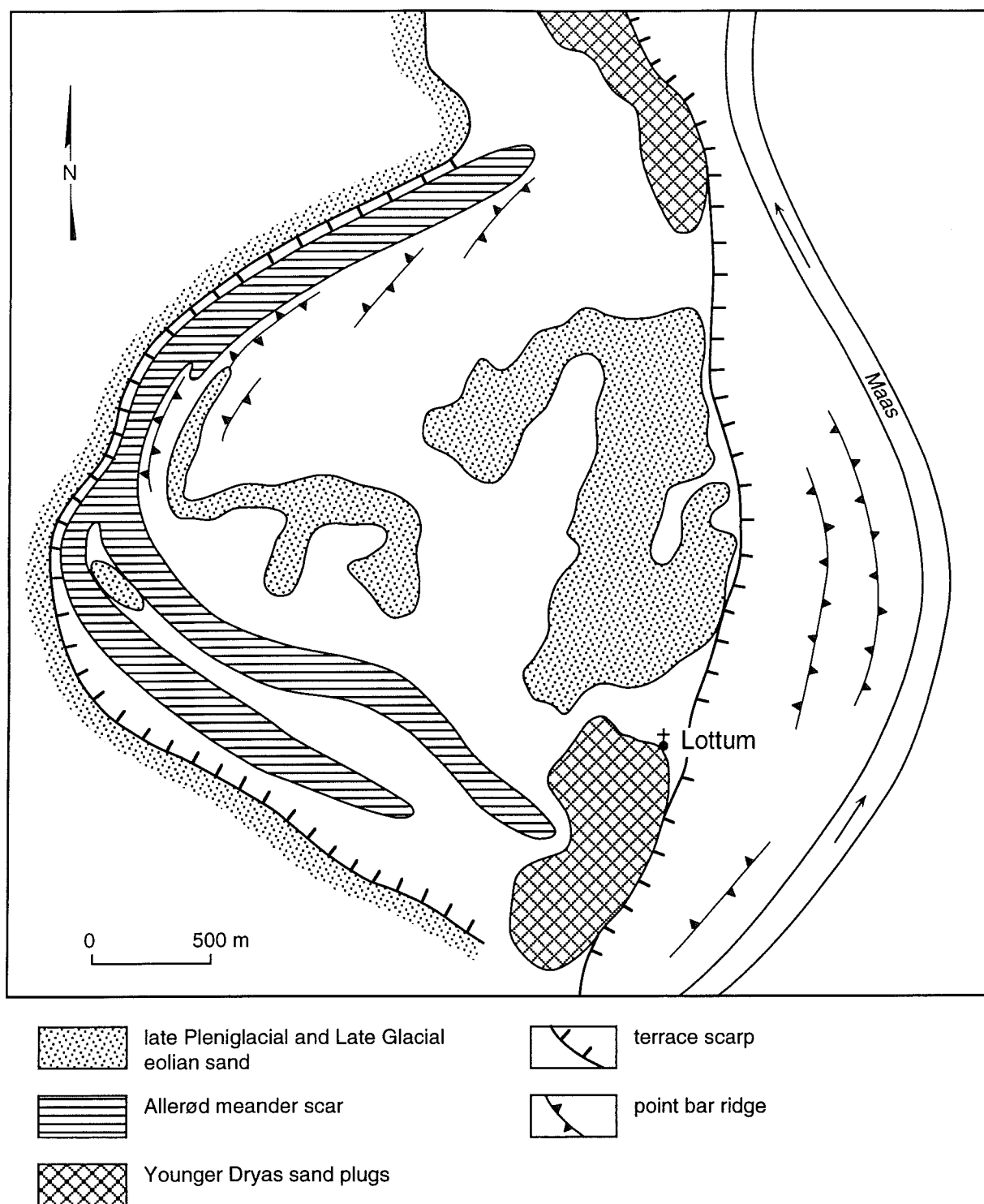


Fig. 4 Younger Dryas chute cut-off of the Allerød meander at Lottum (from Kasse, in press).

Geomorphology and fluvial development

A generalised cross section of the Kasteelweg-Ooijen area is given in figure 5. Four physiographic areas/units are distinguished:

1. The high-sinuosity meander scar terrace in the west. It consists of gravelly sands fining upward into sand and it dates from the early Late Glacial to the Allerød period.
2. The Younger Dryas channel system, underlain by gravelly sands. Local organic channel fill dates from the late Younger Dryas (Westerhoff & Broertjes, 1990: pollen diagram Blitterswijk Linkstraat; Fig. 6: pollen diagram Kasteelweg).
3. The Younger Dryas channel and bar system, partly covered by river dunes, and resting on erosional remnants of the meander scar terrace.
4. The Early Holocene floodplain with channels filled with coarse-detrital gyttja, dating from the late Boreal-early Atlantic (Figs. 7 and 8: pollen diagrams Ooijen).

The first stop is in a 500 m broad, but shallow, Younger Dryas channel (Fig. 5: Kasteelweg). Locally, a 2 m thick organic fill was found in scours within the channel. Palynological research (see discussion on pollen diagram Kasteelweg, Fig. 6) indicates that the base of the infilling (= moment of abandonment of the channel) dates from the end of the Younger Dryas. This age proves that the channel was in use during the Younger Dryas itself.

Walking towards the northeast, physiographic unit 3 is crossed. It is a higher area of straight gravelly channels and interchannel bars, covered by aeolian sand. There is no organic material in these higher lying channels, but on morphological grounds, this channel-bar system is regarded as part of the Younger Dryas floodplain. The dunes besides the channels are interpreted as Younger Dryas river dunes, which were formed by deflation from the neighbouring channels during low water discharges. Corings in area 3 revealed the existence of older Late Glacial meander sediments in the subsurface.

The high topographic position of area 3 (the Younger Dryas channel and bar system) gives the impression of an island, surrounded by lower-lying channels of physiographic units 2 and 4. This might indicate that erosion occurred during the Younger Dryas period: the Younger Dryas channels, which first flowed on the topographical level of area 3, gradually incised towards their position in area 2 and probably also area 4.

At the east side of physiographic unit 3 a clear morphological boundary is present between the Younger Dryas dunes and physiographic unit 4: the Holocene floodplain. This boundary is an erosional one, made by a Holocene channel which flowed east of the Younger Dryas dune field. After channel abandonment it was filled with up to 3.5 m of clay and coarse-detrital gyttja (Figs. 7 and 8: pollen diagrams Ooijen). Several detailed cross-sections were made over the channel. Generally, the fill is 3 m thick and registration started in the Atlantic (Fig. 8). Only very locally the fill is thicker (3.5 m) and registration started already during the late Boreal (Fig. 7), which stresses the importance of a detailed survey.

SW

PHYSIOGRAPHIC UNITS

NE

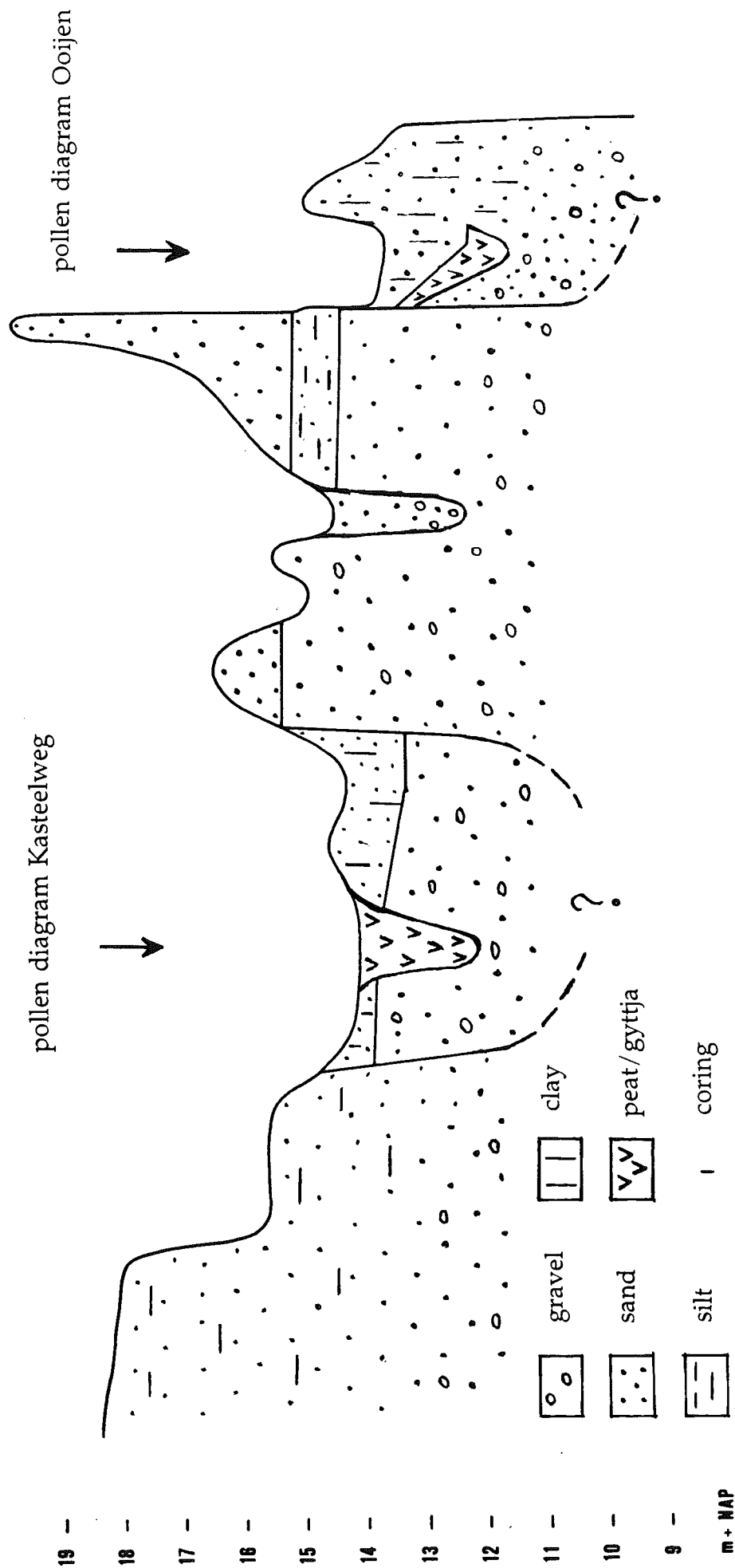


Fig. 5 Schematic cross section and physiographic units of the Younger Dryas and Holocene floodplains at Kasteelweg-Ooijen.

The palynology of the fill is discussed below. Since the base of the fill dates from the late Boreal it is clear that the erosion of the Younger Dryas dunes occurred during the Preboreal or early Boreal. Some small finds of Mesolithic flint artifacts on top of the Younger Dryas dunes point to occupation along the erosional boundary and the abandoned channel.

The base of the abandoned early Holocene channel (Fig. 5: Ooijen) is situated at almost the same level as, or slightly lower than, the base of the Younger Dryas channel (Fig. 5: Kasteelweg). This possibly indicates that at the end of the Younger Dryas two channel systems existed, west and east of physiographic area 3. The climatic transition towards the Preboreal probably resulted in a more regular and/or lower discharge, which led to the abandonment of the western branch (Kasteelweg) of the Younger Dryas system. By this change from a multi-channel system into a single-channel system the Ooijen channel became the principal river course at the start of the Holocene.

The pollenrecord at Kasteelweg (Fig. 6)

Organogenic deposits in the straight braided channels of Younger Dryas age are rare, but within the limits of the former wide stream-bed in some depressions organics have been preserved. At Kasteelweg a black gyttja going into a greenish gyttja was found underlying the organic sediments that were more frequently encountered. At the transition between these two lithological units a 1 mm thin sand band and a thin gray clay layer are present (see lithological column Fig. 6).

Local pollenzone KAW-1 (193-167 cm):

This zone is characterized by the presence of Gramineae, *Artemisia*, *Thalictrum* and *Betula*. *Betula* shows an increase towards the top of this zone (35%). *Juniperus* is only rarely present. Since we are on the Younger Dryas terrace and the overall picture of this zone is one that is dominated by the NAP, the pollenrecord is thought to demonstrate the termination of the **Late Dryas biozone**. The increase in *Betula* towards the top of this zone is interpreted as the Early Holocene *Betula*-rise. The absence of any reaction at this level in the *Juniperus* curve remains peculiar. It can not be excluded that the thin sand band, forming the transition to the overlying zone, reflects a truncation of the underlying zone and that the early Holocene is represented by a hiatus in the sequence.

Local pollenzone KAW-2 (167-143 cm):

After a possible hiatus at the level of the thin sand band registration resumes, while clastic sedimentation takes place. Reworked taxa occur at this level (*Alnus*, *Corylus*, *Quercus*, *Acer*). *Betula*, Gramineae and *Artemisia* maintain in the lower part of this zone, but subsequently the curves of these taxa show a declining tendency and *Thalictrum* is absent from the pollenrecord. Remarkable is the appearance of *Populus* directly after the supposed erosional hiatus.

Voor barasing
DEWOUT
baring

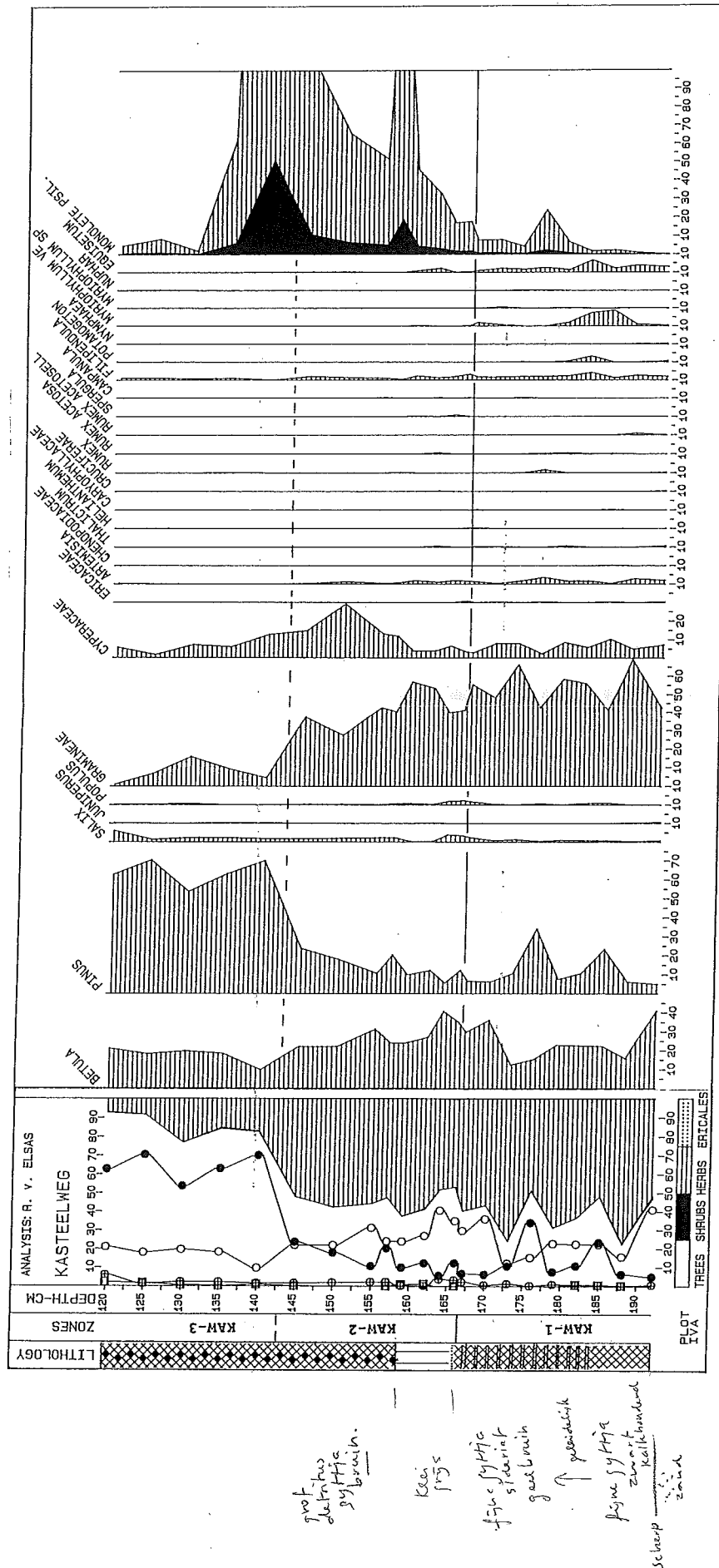


Fig. 6 Pollen record (selection of curves only) from the infill of the Younger Dryas braided channels at Kasteelweg.

The presence of *Populus* at this level in the sequence may possibly indicate that the vegetational record can be placed in the **Rammelbeek-phase** of the **Preboreal**.

The upper half of this zone is mainly characterized by a dominance in the Cyperaceae. The Cyperaceae peak is thought to be a local effect of an hydrosere succession following the deposition of the clay.

Local pollenzone KAW-3 (143-120 cm):

The sharp pine increase, indicative for the spread of *Pinus* in the region of the Maas valley, is dated to 9,700 BP in a meander scar due south of this location (Lottum-Schuitwater). *Corylus* only shows a slight increase in the uppermost samples of the analyzed core segment and indicates the end of the **Preboreal** period (c. 9150 BP).

Resuming, the pollenrecord shows the termination of the **Late Dryas**, a hiatus towards the early Holocene and the later part of the **Preboreal**. At the start of the **Preboreal** (c. 10,200 BP) or shortly after and before 9,700 BP, river erosion and subsequent deposition of clay from suspension load took place originating from the then active early Holocene river system.

The Holocene flood plain at Ooijen

Two cores were taken in the early Holocene river plain at Ooijen (Figs. 1, 3 and 5) in order to establish the period during which the channel became abandoned and started to fill in with organogenic sediments (Figs. 7 and 8).

Core 91-12 reached a greater depth than core 91-11 and also on lithological grounds appeared to contain a different type of sediment in its basal part. Here a coarse-detrital clay was present below the grey sandy clay that was encountered everywhere in the channel fill.

Local pollenzone OOIJ-1 (348-320 cm, core 12):

This zone represents a *Pinus* dominated phase with pine values up to 80%. *Corylus* remains relatively low and *Ulmus* and *Quercus* occur in low frequencies. The pollen assemblage agrees well with the **early Boreal** vegetational history, where within a mature boreal pine forest *Corylus* gradually spread.

Local pollen zone OOIJ-2 (320-283 cm, core 12):

During this zone *Corylus* shows a gradual rise followed by a spread of the Quercetum mixtum (*Quercus*, *Ulmus*, *Tilia*, *Fraxinus*). This succession is typical for the upper half of the **Boreal** period from c. 8400 BP. The spike in the *Pinus* curve at 284 cm does fit the picture for the late **Boreal** where a last *Pinus* maximum is generally present in the pollen records. An estimated date for this dry interval comes to 8250 - 8050 BP (Bohncke, 1991).

Local pollenzone OOIJ-3 (283-281 cm, core 12; 292-256 cm, core 11):

The *Alnus* increase in the top most sample of core 12 indicates that the vegetational record approaches the transition to the **Atlantic** period, c. 8000 BP.

Local pollenzone OOIJ-4 (256-130 cm, core 11):

Core 11 demonstrates that after a gradual increase in *Alnus* a rapid expansion of the species takes place resulting in a relative decline in *Corylus* and *Pinus*. The *Pinus-Alnus* crossing in the pollenrecord is taken to represent the start of the **Atlantic** period. It can not be excluded that at 217 cm depth an erosion hiatus is present in the sequence. Both the lithology (the occurrence of clay pebbles) and the sudden decline in both *Tilia* and *Ulmus* may be taken as indicative for such a feature. Further research may elucidate this problem.

Resuming, the biostratigraphical indications as provided by the pollen records from core 11 and 12 demonstrate that the infilling of the channel at Ooijen started shortly after the **Preboreal/Boreal** transition. Accumulation of clayey detrital gyttja and clay continued over the **Boreal** and **early Atlantic** period and probably even longer.

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EXCURSION STOP 5: ALLERØD MEANDER SCAR AT BEUGEN

Introduction

Beugen is located on the left bank of the Maas, 20 km southeast of Nijmegen (Fig. 1). At this excursion point a well-developed Late Glacial meander scar is present, previously investigated by Teunissen (1990, pollen diagram Helbroek). This scar is situated on the Late Glacial terrace, 11-13 m above sea level (Fig. 2). To the south this terrace with large meander scars is separated from the Pleniglacial floodplain by a terrace edge. At the excursion point Beugen, the Late Glacial meander occurs at nearly the same topographic level as the Pleniglacial floodplain in the west (see Fig. 2). However, erosion did occur during the transition from the braided Late Pleniglacial/Early Late Glacial phase into the Allerød meandering phase (Kasse et al., 1994). Corings in the meander pointbar showed that coarse-grained channel sediments occur at a higher level in the pointbars than in the base of the channel at the moment of cut off. This means that during lateral migration of the channel also vertical erosion took place.

Infill of the meander scar

The meander scar gives a good impression of the Late Glacial Maas system. The scar is approximately 200 m wide and the infilling is up to 3 m thick. The scar is a fine example of a neck cut off. Such neck cut offs are caused by the meandering process itself and they are not related to climatic changes, in contrast to chute cut offs which can be caused by climatic changes as well. Therefore, the start of the infilling of the cut off channel gives a good date of the period in which the Maas floodplain was characterized by large-scale meanders. The fill was studied by several detailed cross sections (Fig. 3). The fill consists of laminated, fine-detrital gyttja at the base (Fig. 3), abruptly lying on the coarse-grained channel sediment deposited before the moment of cut off (coarse sand and gravel). Palynological analyses (see Fig. 4) point to an Allerød age (*Betula* phase) for the start of the infilling. This date shows that the Maas was a large-scale meandering river at least during the Allerød (moment of neck cut off).

The gyttja is erosively overlain by a gray, fine sandy clay, often calcareous at the base (Fig. 3). This unit represents a phase a renewed fluvial activity in the meander scar. Current velocity in the channel was low, however, (deposition of sandy clay) and the Allerød channel morphology was not modified. The erosional transition from gyttja to clay correlates pollen analytically with the Allerød - Younger Dryas boundary. This means that fluvial inundations with clay deposition reached the channel again after a period of organic deposition. The Younger Dryas cooling is held responsible for this change in fluvial dynamics. The decreased evapotranspiration and probably also the increased snowfall in the higher regions of the Maas catchment (Ardennes) resulted in higher discharges and higher peak discharges. These higher

peak discharges are the reason for the renewed inundation of the Allerød meander scar at Beugen and the Late Pleniglacial floodplain at Bosscherheide (excursion stop 2).

The upper two units of the scar infill (Fig. 3) are probably of Holocene age. The sand layer was probably deposited by a local brook (Oeffeltsche Raam), coming from the southwest and flowing through the meander scar. The peat bed (Fig. 3), commonly a wood peat, is rather thin here. Locally, it is thicker and the base dates from the Late Boreal or Early Atlantic (Teunissen, 1990). The peat overlies the Younger Dryas clay, which indicates a hiatus due to non-deposition during the Preboreal and Boreal.

Pollenrecord of the infill

The results of the pollen analyses are presented in Fig. 4. Four local pollen zones have been distinguished.

Local pollenzone BGN-1 (333-322 cm):

The basal spectra of the infill show part of the *Betula*-phase of the Allerød.

Local pollenzone BGN-2 (322-297 cm):

At 321 cm *Pinus* starts to rise. A date for this event at Bosscherheide (Excursion stop 2) comes to 11,300 BP. The *Pinus*-phase of the Allerød is relatively short and it cannot be excluded that during the overlying Late Dryas, part of the Allerød has been truncated. Lithologically a sharp transition from laminated gyttja to slightly sandy clay has been observed at 295 cm depth.

Local pollenzone BGN-3 (297-279 cm):

At 296 cm the *Pinus* value drops, the NAP pollen rises, while the pollen of tree birches show an increase. This short-lasting birch phase at the beginning of the Late Dryas is a repetitive feature in many diagrams of the Maas. At Bosscherheide remnants of birch trees have been revealed from the Late Dryas organic deposits in the backswamp, that were formed shortly after 10,880 BP.

Local pollenzone BGN-4 (279-170 cm):

From 279 cm depth the landscape gradually became more open and grasses, sedges and *Artemisia* together with dwarf birches and dwarf willows and some *Dryas*, *Helianthemum* and *Thalictrum* dominate the picture. The relatively high values for *Classopollis*, a pollen-type derived from the Lower Cretaceous beds in the upper course of the Maas, clearly demonstrates that the ox-bow lake was again reached by the then active Maas river. Probably during periods of peak discharge these old streambeds were carrying water and deposition of suspension load took place.

A transition to the early Holocene has not been reached in the analyses and is probably absent due to Holocene brook incision in the ox-bow infill.



Fig. 1 Location map of the Allerød meander scar at Beugen.

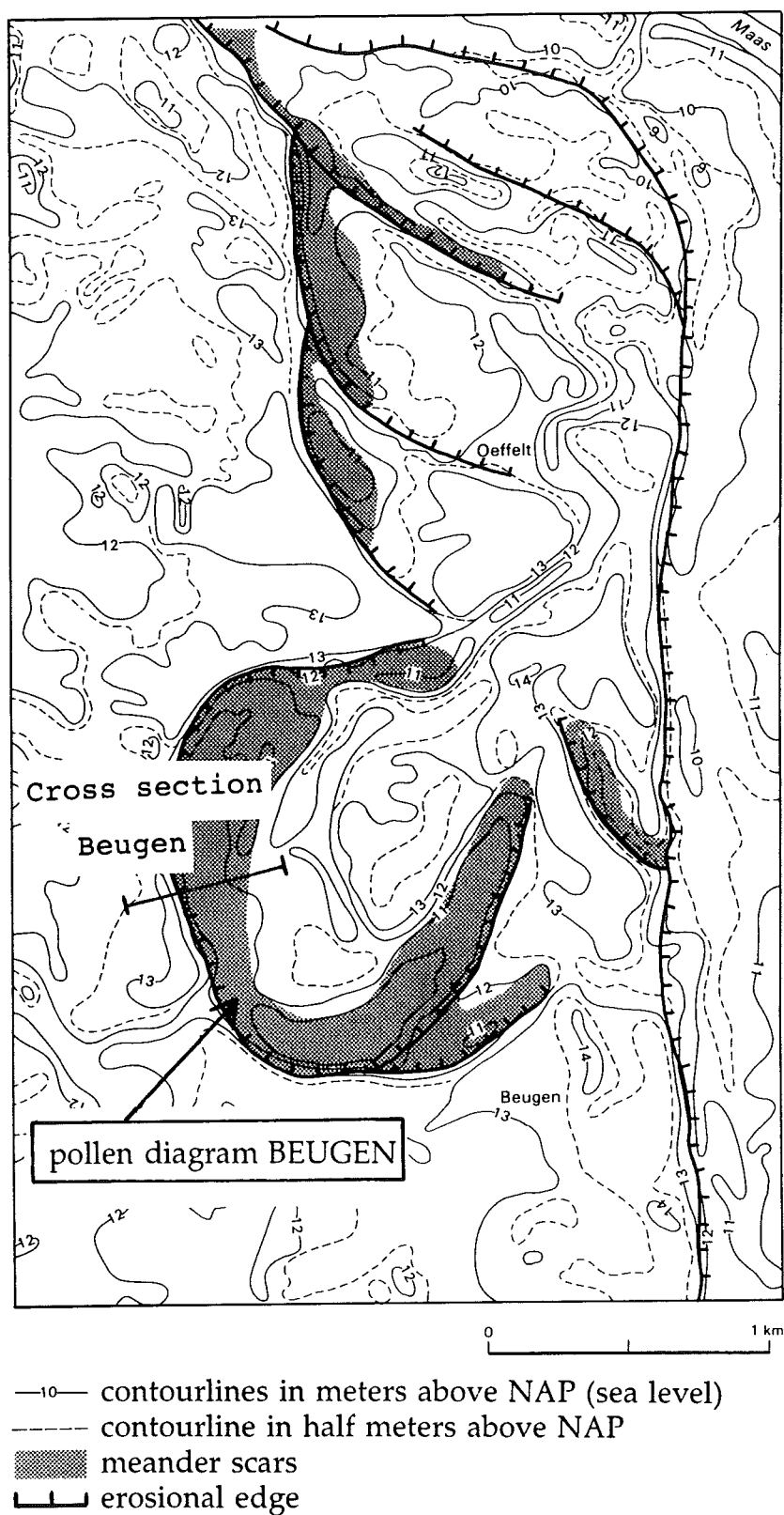


Fig. 2 Morphological map of the Allerød meander neck cut off at Beugen (after Buitenhuis & Wolfert, 1988).

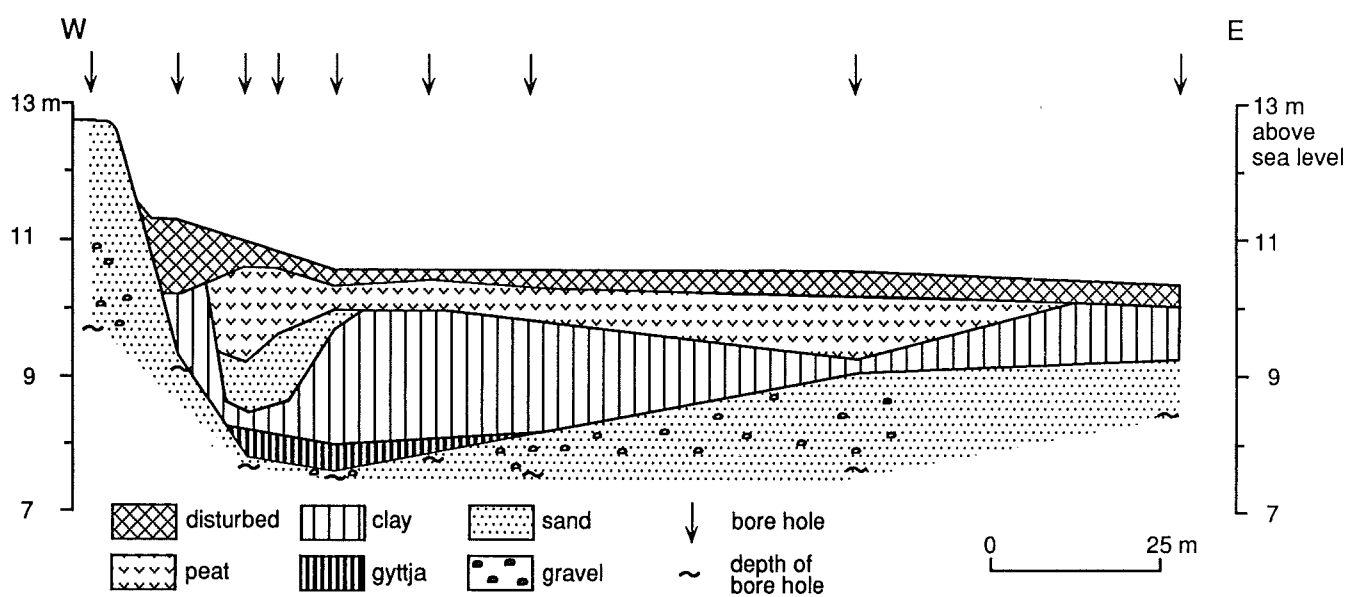


Fig. 3 Cross section over the Late Glacial and Holocene infill of the meander neck cut off at Beugen (from Kasse et al., 1994).

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